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Uncertainty in global groundwater storage estimates in a Total Groundwater Stress framework

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Key Points:

- Groundwater resilience is defined and quantified with remote sensing from GRACE
- Timescales of aquifer depletion are assessed as a Total Groundwater Stress ratio
- The volume of usable global groundwater storage is found to be largely unknown

Supporting Information:

- Supporting Information S1
- Tables S1
- Tables S2
- Tables S3
- Tables S4

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Abstract Groundwater is a finite resource under continuous external pressures. Current unsustainable groundwater use threatens the resilience of aquifer systems and their ability to provide a long-term water source. Groundwater storage is considered to be a factor of groundwater resilience, although the extent to which resilience can be maintained has yet to be explored in depth. In this study, we assess the limit of groundwater resilience in the world's largest groundwater systems with remote sensing observations. The Total Groundwater Stress (TGS) ratio, defined as the ratio of total storage to the groundwater depletion rate, is used to explore the timescales to depletion in the world's largest aquifer systems and associated groundwater buffer capacity. We find that the current state of knowledge of large-scale groundwater storage has uncertainty ranges across orders of magnitude that severely limit the characterization of resilience in the study aquifers. Additionally, we show that groundwater availability, traditionally defined as recharge and redefined in this study as total storage, can alter the systems that are considered to be stressed versus unstressed. We find that remote sensing observations from NASA's Gravity Recovery and Climate Experiment can assist in providing such information at the scale of a whole aquifer. For example, we demonstrate that a groundwater depletion rate in the Northwest Sahara Aquifer System of 2.69 $\pm$ 0.8 km^3/yr would result in the aquifer being depleted to 90% of its total storage in as few as 50 years given an initial storage estimate of 70 km^3.

1. Introduction

Changes in Earth's climate and the increased use of groundwater resources to meet global water demands [*Kundzewicz and Döll*, 2009; *Famiglietti*, 2014] restrict efforts to sustainably govern the common pool resource [*Schlager et al*., 1994; *Dietz et al*., 2003; *Steward et al*., 2009]. Traditionally studied by estimating fluxes, it is now widely recognized that both groundwater fluxes (i.e., discharge and recharge) and stocks (i.e., storage volumes) are necessary to effectively monitor the state of groundwater [*Schlager et al*., 1994; *Steward et al*., 2009]. Globally, large uncertainty in global groundwater storage exists [*Alley*, 2006] limiting our ability to characterize groundwater resilience, a function of aquifer storage, as perturbations to the aquifer result from climatic and anthropogenic influences. The insufficient knowledge of total groundwater supplies will continue to limit effective governance of groundwater systems until a significant effort is made to improve groundwater storage estimates.

Groundwater storage estimates commonly cited in global groundwater assessments [*Graham et al*., 2010; *Perlman*, 2012] can be traced to decades-old studies [*Korzun*, 1974, 1978; *Baumgartner and Reichel*, 1975; *Nace*, 1969; *Lvovich*, 1974; *Berner and Berner*, 1987]. The historical estimates, however, vary from 7 $\times$ 10^6 cubic kilometers (km^3) to 23 $\times$ 10^6 km^3 [*UN World Water Assessment Program*, 2003]. These largely uncertain values have filtered into the global groundwater literature, and although they were derived only heuristically, they have become commonly accepted. For example, although there is no observational basis, it is commonly accepted that groundwater comprises 30% of global freshwater [*Shiklomanov*, 1993] as calculated using the upper estimate of global groundwater storage by *Korzun* [1978].

Hashimotoet al. [1982] put forth the concept of quantifying resilience of water resource systems with an application to a water supply reservoir. *Sharma and Sharma* [2006] define groundwater resilience as the "ability of the system to maintain groundwater reserves in spite of major disturbances." Remote sensing of terrestrial water storage changes provides a valuable tool to observe and isolate changes in subsurface water storage that result from disturbances, both natural and anthropogenic, that influence the resilience of groundwater systems.

The concepts of stability and resilience in reference to ecological systems state that a natural system undergoes perturbations from an equilibrium state. Stability accounts for the time to return to normal and resilience accounts for the amount of disturbance while maintaining a state of equilibrium [*Holling*, 1973]. The ability of groundwater to provide resilience is rooted in the large storage capacity of groundwater systems [*Anderies et al*., 2006; *Sharma and Sharma*, 2006; *Shah*, 2009; *MacDonald et al*., 2011; *Hugman et al*., 2012; *Katic and Grafton*, 2011; *Taylor et al*., 2013] and the residence time of groundwater, typically orders of magnitude larger than residence times for surface water [*MacDonald et al*., 2011; *Lapworth et al*., 2012]. However, additional negative impacts of pumping can significantly reduce the resilience of systems even with large volumes of water in storage. These factors include the influence of pumping on subsidence, streamflow depletion, increased drilling and pumping costs, decreasing water quality with increased use, and the volume actually recoverable from an aquifer, which is less than total storage [*Alley*, 2007]. A number of studies have assessed groundwater resilience during drought [*Peters et al*., 2005; *Hugman et al*., 2012] or resulting from hydroclimatic variability [*Lapworth et al*., 2012].

We apply the definition of resilience from *Holling* [1973] and *Sharma and Sharma* [2006] to groundwater such that the resilience of a coupled human-groundwater system is the ability of the system to increase recharge and decrease base flow, termed "capture" [*Lohman*, 1972], to maintain equilibrium or to reach a new equilibrium as suggested by *Theis* [1940]. When equilibrium is unattainable, at least over the timescales of interest, there is a loss of groundwater storage [*Theis*, 1940; *Alley et al*., 2002; *Alley and Leake*, 2004] identified as a tipping point for when the limit of the system's resilience is surpassed.

Numerous groundwater studies have shown that groundwater is being used at rates that exceed natural rates of recharge globally [*Döll*, 2009; *Wada et al*., 2010; *Gleeson et al*., 2012; *Richey et al*., 2015]. The importance of groundwater resilience lies in the fact that groundwater is a coupled human-natural system [*Steward et al*., 2009] providing critical services to human and natural ecosystems. Its ability to do so indefinitely relies on the balance between the volume of water that enters a groundwater system and the volume that leaves the system. In a natural system and over long time periods, the average input (i.e., recharge) is balanced by average output (i.e., base flow and evapotranspiration).

Groundwater pumping is a perturbation to the natural system that disrupts the long-term equilibrium state [*Theis*, 1940; *Bredehoeft et al*., 1982; *Alley et al*., 2002; *Mays*, 2013; *Steward et al*., 2013]. Such a perturbation can require tens to hundreds of years to work through the system and re-establish equilibrium, if equilibrium is attainable at all [*Bredehoeft et al*., 1982; *Alley et al*., 2002; *Alley and Leake*, 2004]. The combination of hydroclimatic driven variations from steady state, including the potential influence of climate change [*Döll*, 2009; *Taylor et al*., 2013], and human perturbations from agricultural and urban development [*Alley et al*., 1999; *Alley and Leake*, 2004; *Mays*, 2013] will continue to produce deviations from steady state in groundwater systems. These deviations could lead to long-term depletion of groundwater storage on human timescales [*Mays*, 2013] resulting in reduced resilience and increased groundwater stress. The resilience of an aquifer system against unsustainable pumping can be improved with human intervention, but only where transparent knowledge of the system exists [*Liu et al*., 2007; *Folke et al*., 2002]. This study increases the transparency of knowledge on the state of fluxes and stocks in large aquifer systems to assess aquifer resilience.

NASA's Gravity and Recovery and Climate Experiment (GRACE) satellite provides vertically and spatially integrated observations of changes in snow, soil water, surface water, and groundwater storage over a region [*Tapley et al*., 2004]. Using auxiliary data sets, we can use GRACE observations to provide the first-ever observation-based quantification of groundwater resilience whereby subsurface storage changes are combined with aquifer storage estimates to assess the limits of groundwater resilience. The goals of this study are threefold: (1) to summarize and evaluate estimates of global groundwater storage by characterizing the large variability of current estimates using traditional hydrogeologic characteristics; (2) to evaluate groundwater resilience as a function of global groundwater storage estimates and the net average subsurface storage flux as observed using remote sensing; and (3) to highlight how the current, uncertain knowledge of global groundwater storage severely undermines efforts to quantify resilience. We conduct our evaluation on the 37 "Large Aquifer Systems of the World" [*WHYMAP and Margat*, 2008] (Figure 1). Simplifying assumptions are made to the study due to limited data availability and consistency across the study aquifers.

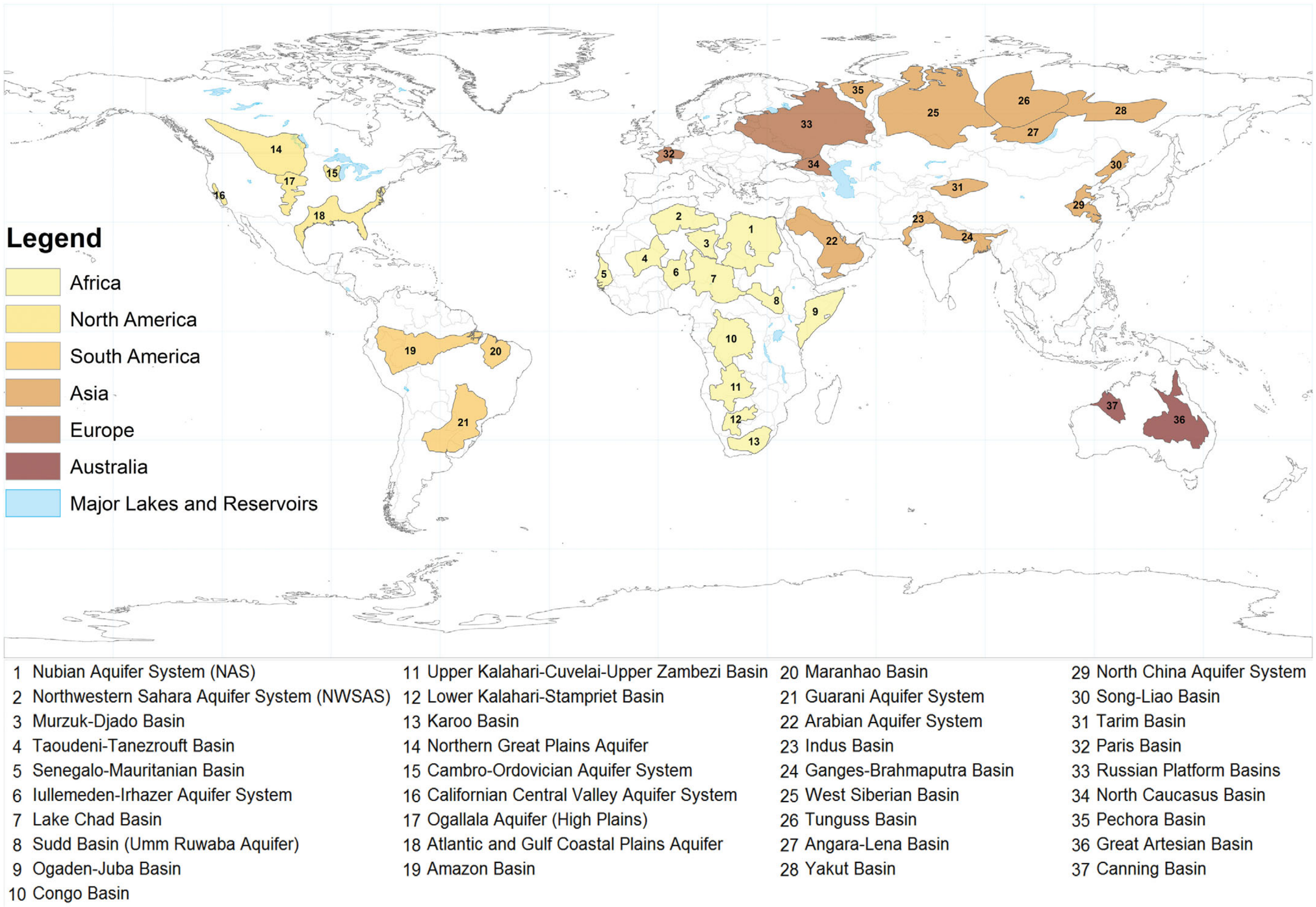

Figure 1. Study aquifers by continent based on the WHYMAP delineations of the world's large aquifer systems [*WHYMAP and Margat*, 2008]. The number represents the aquifer identification number associated with the aquifer name listed below for each system. The world's largest lakes and reservoirs are based on the Global Lake and Wetland Database Level-1 lakes and reservoirs [*Lehner and Döll*, 2004].

2. Data and Methods

Previous water stress studies [*Alcamo et al.*, 1997; *Vörösmarty et al.*, 2000; *Oki and Kanae*, 2006; *Döll*, 2009; *Richey et al.*, 2015] evaluated the ratio of water use to water availability, where groundwater availability was defined as the volume of recharge to the aquifer system. Such approaches highlight historical groundwater use in relation to renewable groundwater recharge to manage groundwater systems yet fail to account for the buffer capacity of aquifer storage. For example, an aquifer system with little storage has a limited ability to buffer against drought and excessive groundwater pumping as compared to an aquifer system with large storage volumes.

2.1. Subsurface Depletion

We use observations from the Gravity Recovery and Climate Experiment (GRACE) satellite mission [*Tapley et al.*, 2004] to quantify changes in subsurface storage in the study aquifers. GRACE is a joint mission between the United States and the Deutsche Forschungsanstalt für Luft und Raumfahrt (DLR) in Germany to monitor changes in Earth's gravity field that can be used to isolate time-variable anomalies in terrestrial water storage. The Center for Space Research at the University of Texas at Austin provided the 132 months of GRACE gravity coefficients from Release-05 data used in this study.

Gravity anomalies from GRACE observations are processed for the study period (January 2003 to December 2013) to produce average terrestrial water storage anomalies for each of the 37 study aquifers [*Swenson and Wahr*, 2002; *Wahr et al.*, 2006; *Swenson and Wahr*, 2006]. The processing results in some lost signal power from truncating the gravity coefficients (at degree and order 60) and filtering. Aquifer-specific scaling factors are used to account for the lost signal power and to estimate unbiased mass change in each aquifer system [*Velicogna and Wahr*, 2006]. The truncation has a greater influence on smaller regions, therefore accuracy of GRACE estimates increases as the area of the region of interest increases [*Wahr et al.*, 2006].

Total water storage anomalies are isolated from the total gravity anomalies as the time-variable component of the GRACE signal, representing combined natural (*N*), and anthropogenic (*A*) anomalies in snow water equivalent (*SWE*), soil moisture (*SM*), groundwater (*GW*), and surface water (*SW*) according to equation (1). Anomalies in individual storage components can be isolated by removing anomalies in the remaining storage terms from ΔS_{N+A} with independent estimates of these components according to equation (2) [*Rodell and Famiglietti*, 2002; *Swenson et al.*, 2006; *Yeh et al.*, 2006; *Strassberg et al.*, 2007, 2009; *Rodell et al.*, 2004b, 2007, 2009; *Swenson et al.*, 2008; *Famiglietti et al.*, 2011; *Scanlon et al.*, 2012]. In equation (2), we isolate subsurface storage, defined as the combination of soil moisture and groundwater, to be consistent with total storage estimates that begin at the ground surface as defined in section 2.2. Auxiliary data sets for *SWE* and *SW* are necessary to separate subsurface anomalies for the remaining storage terms.

$$\Delta S_{N+A} = \Delta(SW + SWE + SM + GW)_{N+A} \tag{1}$$

In our evaluation, we use model output from the NASA Global Land Data Assimilation System (GLDAS) [*Rodell et al.*, 2004a] including the Noah [*Chen et al.*, 1996; *Koren et al.*, 1999], Community Land Model 2.0 (CLM2) [*Dai et al.*, 2003], and Variable Infiltration Capacity (VIC) [*Liang et al.*, 1994] models to quantify natural changes in *SWE* and canopy surface water (*CAN*). Surface water storage (*SW*) as stocks in lakes, reservoirs, and river channels is not included in the GLDAS modeling system; thus, for *SW*, we use the sum of routed surface water discharges (*RIV*) from the Community Land Model 4.0 (CLM 4.0) [*Oleson et al.*, 2010] and *CAN*. CLM 4.0 was driven in an offline simulation by 3 hourly precipitation, near surface air temperature, solar radiation, specific humidity, wind speed, and air pressure from GLDAS Version-1 [*Rodell et al.*, 2004a]. The model was run for the study period at 0.9° × 1.25° spatial resolution and linearly interpolated to 1° × 1° and monthly temporal resolution.

$$\Delta SUB_{N+A} = \Delta S_{N+A} - \Delta(SW + SWE)_N \tag{2}$$

Anomalies in subsurface storage are computed as the residual between the GRACE total water storage anomalies and the model-based storage anomalies of *SWE* and *SW* for each study aquifer. We assume the anthropogenic influence on the storage anomalies is dominated by subsurface variations, particularly from groundwater, and that the direct anthropogenic influence on the remaining storage components is negligible in comparison. This is due to the spatial scale of the infrastructure necessary to capture snow and surface water for anthropogenic uses being significantly smaller than the aquifer study areas [*Vörösmarty et al.*, 2000; *Richey and Famiglietti*, 2012].

The subsurface error was calculated using equation (3) for each month (*i*), assuming the errors from each storage component are independent. Aquifer-specific satellite measurement and leakage error from processing the gravity anomalies is computed following *Wahr et al.* [2006] to estimate error in the total GRACE signal. Variance of *SWE* and *CAN* was determined using the three-model ensemble and thus represents a combination of estimate error and model representation error. The U.S. Geological Survey errors for hydrologic measurements range from excellent (5% error) to fair (15% error) [*U.S. Geological Survey*, 2014]; for our evaluation, we assume measurement error of 50% for *RIV* to represent a conservative uncertainty in GRACE subsurface variability.

$$\sigma_{\Delta SUB,i} = \sqrt{\sigma_{S,i}^2 + \sigma_{SWE,i}^2 + \sigma_{CAN,i}^2 + \sigma_{RIV,i}^2} \tag{3}$$

A conservative estimate of groundwater trends can be identified if we attribute observed subsurface trends solely to groundwater storage. Such an approach is considered here, as the majority of soil mass trends are not significant globally [*Sheffield and Wood*, 2008; *Dorigo et al.*, 2012]. We consider the groundwater trend, ΔGW_{trend}, to be representative of the net flux of water storage resulting from groundwater use (ΔGW_{N+A}), including the aquifer response to pumping, and natural climatic variability. Annual trend magnitudes were estimated using the weighted regression in equation (4). The weights, w_i, are a function of the variance in the monthly estimates of subsurface storage anomalies. Aquifers with a negative coefficient were

10.1002/2015WR017351

considered to be depleting in aquifer storage while positive coefficients were considered to be recharging systems. Here, we evaluate only the magnitude of trends without regard to trend significance. Figure 4d from *Richey et al*. [2015] demonstrates this method by showing the time series and associated trend in subsurface storage anomalies for the Ganges-Brahmaputra Basin (Aquifer 24, "Ganges"). A comparison between depletion estimates from this study are compared to available depletion estimates in eight study aquifers (supporting information Text S1).

$$Y_i = \beta_0 + \beta_1 w_i x_i + \varepsilon_i \ \text{ where } \ w_i = \frac{1}{\sigma^2_{\Delta SUB,i}} \tag{4}$$

2.2. Water Availability

Groundwater is frequently pumped beyond the renewable rate thus depleting groundwater storage over time [*Theis*, 1940; *Sahagian et al*., 1994; *Konikow and Kendy*, 2005; Famiglietti, 2014]. This study revises the definition of groundwater availability from recharge, as previously used in a stress framework [*Döll*, 2009; *Wada et al*., 2010; *Richey et al*., 2015], to total groundwater storage, as recommended by *Taylor* [2009]. Defining groundwater availability as the total volume of groundwater in storage allows for the concepts of resiliency and buffer capacity to be explored as a component of groundwater stress. This is important in regions that may be considered stressed based on renewable supplies of groundwater but that contain a large volume of storage [*MacDonald et al*., 2012].

We group available and revised storage estimates into three categories: historical, regional, and revised estimates. The historical estimates distribute the range of most commonly cited global groundwater storage estimates into the study aquifers. The regional estimates are comprised aquifer-specific storage estimates from readily accessible regional case studies. We develop the revised estimates based on modifications to the historical methods to constrain the large uncertainty range in existing aquifer storage estimates. Aquifers with high and low volumes of storage are compared to aquifers with high and low volumes of recharge to show the influence of defining availability based on a renewable flux or total stocks. Basin-averaged mean annual recharge from *Richey et al*. [2015] is used to quantify recharge in this study. Negative values of recharge highlight basins where upward capillary fluxes are the dominant subsurface flux and positive values indicate a downward flux [*Richey et al*., 2015].

2.2.1. Historical Storage Estimates

Nace [1969] and *Korzun* [1978] provide lower and upper total estimates of global groundwater storage, respectively. The storage limits were calculated by equation (5), with the difference between the estimates originating in the respective values used for effective thickness (b) and porosity (n) in the subsurface. The historical approaches assume uniform groundwater supply across the global land area (A), excluding Greenland and Antarctica; however, it is unrealistic to assume uniform groundwater across the global land area [*Alley*, 2006].

$$V = bnA \tag{5}$$

The lower storage boundary (7×10^6 km^3) assumed an effective subsurface thickness as 1000 m and an effective porosity of 1% over the global land area. An arbitrary increase to the resulting volume was applied based on the author's belief that the total storage should be approximately five times greater than the calculated storage [*Nace*, 1969]. The upper boundary (23×10^6 km^3) calculated storage for each continent by dividing the subsurface into three zones of varying thickness with associated porosity of 15%, 12%, and 5% depending on depth (Table 1) totaling 2000 m. Continental groundwater storage was then summed to obtain the global storage estimate. *Shiklomanov* [1993] warned that these estimates are inaccurate approximations based on coarse assumptions.

In the present study, we distribute the global values by *Nace* [1969] and *Korzun* [1978] into the study aquifers based on an area weighted scheme by assuming the majority of global groundwater in storage exists in the largest global aquifers (supporting information Table S1) [*Margat*, 2007; *Margat and van der Gun*, 2013]. The lower global groundwater estimate by *Nace* [1969] (7×10^6 km^3) was distributed into the study aquifers based on a ratio of each aquifer's area to the total global aquifer area. The upper global groundwater estimate by *Korzun* [1978] (23×10^6 km^3) first calculated continental groundwater storage with

Table 1. Adapted From *Korzun* [1978] as Inputs to Equation (5) to Provide the Upper Bound of the Historical Estimate of Storage

Continent	Total Area (10^6 km^2)	Thickness of Zone (m)	Effective Porosity (%)	Groundwater Volume by Zone (10^6 km^3)	Groundwater Volume by Continent (10^6 km^3)
Europe	10.5	100	15	0.2	1.6
		200	12	0.3	
		2000	5	1.1	
Asia	43.5	200	15	1.3	7.8
		400	12	2.1	
		2000	5	4.4	
Africa	30.1	200	15	1	5.5
		400	12	1.5	
		2000	5	3	
North America	24.2	200	15	0.7	4.3
		400	12	1.2	
		2000	5	2.4	
South America	17.8	100	15	0.3	3
		400	12	0.9	
		2000	5	1.8	
Australia and Oceania	8.9	100	15	0.1	1.2
		200	12	0.2	
		2000	5	0.9	
				Total	23.4

varying saturated thickness by continent. Therefore, the continental storage estimates (Table 1) are individually distributed based on the ratio of each aquifer's area to total aquifer area for each continent.

2.2.2. Regional Storage Estimates

Where available, regional estimates of storage for individual study aquifers are used from a combination of regional case studies [*Al-Ibrahim*, 1991; *Llamas et al.*, 1992; *Swezey*, 1999; *Wallin et al.*, 2005; *Tujchneider et al.*, 2007] and compilations of regional data [*Sahagian et al.*, 1994; *Vrba and van der Gun*, 2004; *Margat and van der Gun*, 2013; *MacDonald et al.*, 2012] as summarized in supporting information Table S1. The historical and revised estimates are compared to the regional estimates, when available. The regional estimates are most commonly derived from groundwater models [e.g., *Cao et al.*, 2013] or measured estimates of saturated thickness and aquifer parameters [e.g., *Williamson et al.*, 1989]. Therefore, the regional estimates provide both an initial constraint on the range of possible storage and summarize the state of knowledge on aquifer-specific storage estimates.

Additional processing was applied to the original storage estimates by *MacDonald et al.* [2012]. They provide a gridded range of groundwater storage estimates across Africa at five kilometer spatial resolution. These estimates were linearly interpolated to $1° \times 1°$ resolution and our 13 study aquifers in Africa were isolated from the available data. We determine total storage in the study aquifers based on the minimum, mean, and maximum values across the range of storage estimates.

2.2.3. Revised Storage Estimates

We implement two revisions to the approaches of *Nace* [1969] and *Korzun* [1978] to constrain aquifer storage estimates using hydrogeologic assumptions. The first constraint replaces the porosity in equation (5) with specific yield (S_y) to represent the volume of groundwater that is extractable from storage [*Johnson*, 1967; *Williamson et al.*, 1989; *Alley*, 2006]. Using specific yield instead of porosity reduces the estimate of storage, since not all water in the pore space is extractable as is indicated by the use of porosity [*Alley*, 2006]. To estimate specific yield, we apply a $1° \times 1°$ global, USDA soil texture class map that was derived for GLDAS [*Rodell et al.*, 2004a] based on percentages of sand, silt, clay from the soils data set of *Reynolds et al.* [2000] (Figure 2a). The most common soil class in each aquifer is determined as the mode for each aquifer. We then overlay the simplified soil classification triangle in Figure 2b [*International Labour Organization (ILO)*, 1987] with the specific yield triangle in Figure 2c [*Johnson*, 1967]. The soil classification and specific yield triangles are used to determine a range of specific yield values (S_y) for each aquifer's dominant soil class to be used in equation (6) to calculate aquifer storage. These values are summarized in Table 2. We follow the suggestion of 200 m as an average limit to the active zone of groundwater exchange to represent saturated thickness in equation (6) along with aquifer area (A_{aq}) [*Margat and van der Gun*, 2013].

$$V_{aq} = (200m) S_y A_{aq} \tag{6}$$

$$V_{aq} = b(1\%) A_{aq} \tag{7}$$

The second constraint focuses on the saturated thickness in equation (5). *Nace* [1969] and *Korzun* [1978] used 1000 and 2000 m, respectively, although it is unlikely that the water at these depths is fully accessible and of a high enough quality to use, given that groundwater quality generally decreases with depth

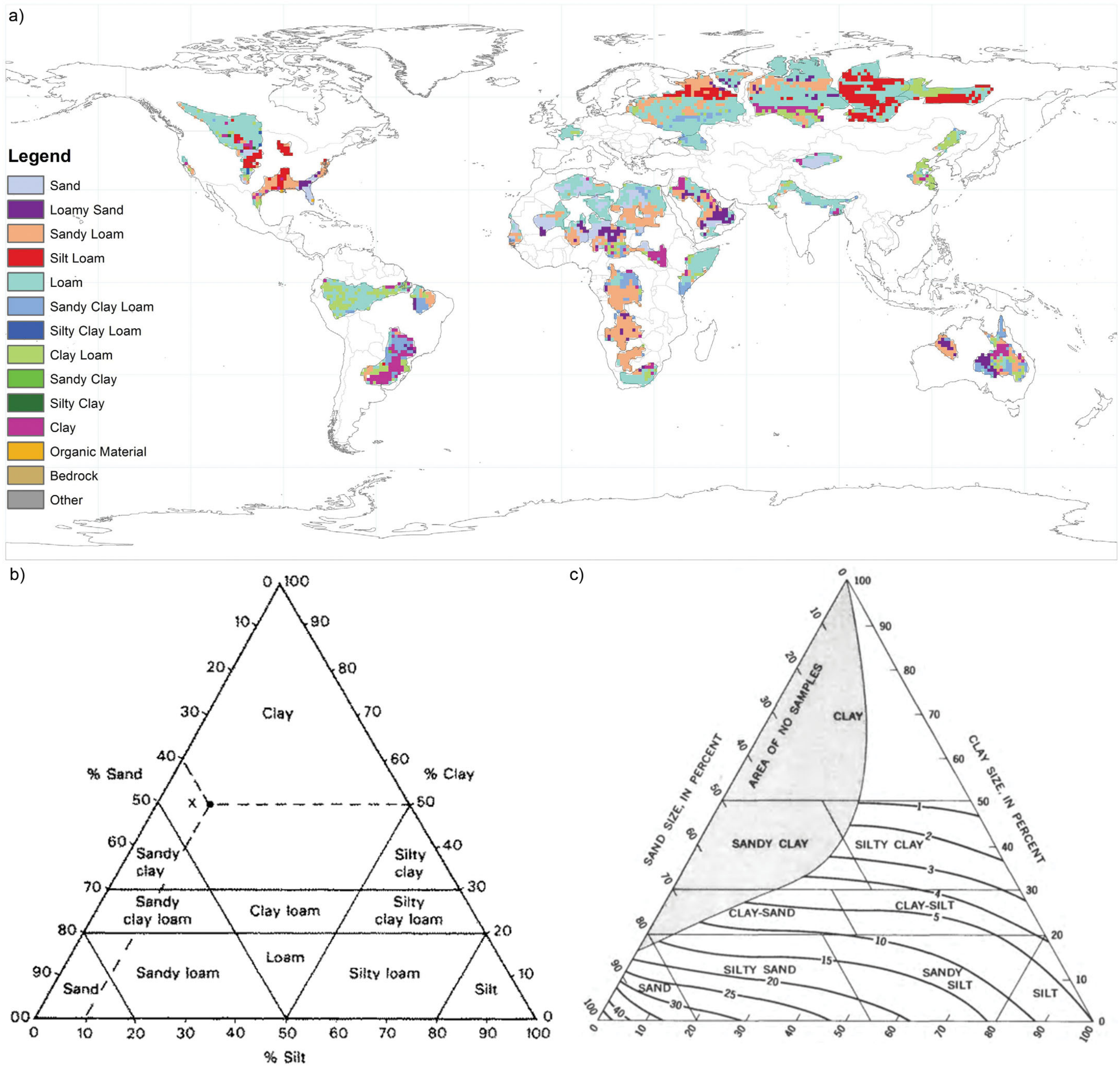

Figure 2. (a) 1° × 1° global-gridded USDA soil texture class map in the study aquifers derived for GLDAS [*Rodell et al*., 2004a] based on percentages of sand, silt, clay from the soils data set of *Reynolds et al*. [2000] used to determine the dominant soil type in each aquifer system. (b) Simplified soil classification triangle based on percentages of sand, silt, and clay [*ILO*, 1987]. (c) Soil classification triangle coupled with estimates of specific yield from *Johnson* [1967] to determine the range of specific yields associated with each soil type. The contours represent specific yield as a percentage.

[*Alley* 2006, 2007; *Faunt*, 2009]. In this analysis, we evaluate a range of aquifer thickness including 20, 50, 100, 200, 500, and 1000 m. We use a constant porosity of 1% following *Nace* [1969] and aquifer area as the remaining inputs to equation (7). This method identifies the potential storage in the aquifer systems, but does not explicitly identify the depth interval across which the saturated thickness is located. Identifying the water table depth combined with the depth to bedrock would further constrain the accessibility of groundwater as a water supply source. However, this is beyond the scope of the current study, which is limited to quantifying total storage.

Table 2. Determination of Specific Yield for the Study Aquifers[a]

Aquifer ID	Soil Type	Minimum Sy	Mean Sy	Maximum Sy
1	Loam	10	16.5	23
2	Loam	10	16.5	23
3	Loam	10	16.5	23
4	Loam	10	16.5	23
5	Sandy loam	10	22.5	35
6	Sandy loam	10	22.5	35
7	Sandy loam	10	22.5	35
8	Clay	0.5	2.5	4.5
9	Loam	10	16.5	23
10	Sandy loam	10	22.5	35
11	Sandy loam	10	22.5	35
12	Sandy loam	10	22.5	35
13	Loam	10	16.5	23
14	Loam	10	16.5	23
15	Silt loam	4	14	24
16	Sandy loam	10	22.5	35
17	Silt loam	4	14	24
18	Sandy loam	10	22.5	35
19	Loam	10	16.5	23
20	Sandy clay Loam	4	8.5	13
21	Clay	0.5	2.5	4.5
22	Loam	10	16.5	23
23	Loam	10	16.5	23
24	Loam	10	16.5	23
25	Loam	10	16.5	23
26	Silt loam	4	14	24
27	Silt loam	4	14	24
28	Silt loam	4	14	24
29	Clay loam	4	7	10
30	Clay loam	4	7	10
31	Sand	15	30	45
32	Loam	10	16.5	23
33	Loam	10	16.5	23
34	Loam	10	16.5	23
35	Loam	10	16.5	23
36	Sandy clay Loam	4	8.5	13
37	Sandy loam	10	22.5	35

[a]The dominant soil type is determined as the mode of the soil type based on Figure 2a. The soil triangles in Figures 2b and 2c are overlaid to determine the minimum and maximum specific yield value for each aquifer. The average of these values is determined as the mean.

2.3. Total Groundwater Stress

We quantify groundwater resilience to explore the impact of natural and anthropogenic disturbance as observed from GRACE and our revised estimates of groundwater storage. We introduce the Total Groundwater Stress (TGS) ratio that estimates the number of years until the aquifer is depleted in equation (8) to evaluate aquifer resilience. The net groundwater storage changes (ΔGW_{trend}), termed groundwater depletion, are quantified with the GRACE-derived groundwater trend from section 2.1. For the purpose of this study, we assume that the rate of depletion will remain constant into the future.

$$TGS_{p\%} = \frac{V_{p\%}}{\Delta GW_{trend}} \tag{8}$$

The total storage estimates from section 2.2 are used to estimate groundwater storage (V) for each study aquifer. It is unrealistic to fully deplete an aquifer system as changes in water quality, accessibility, and soil properties will limit the amount of groundwater that can be extracted [*Alley*, 2007]. To account for the possible range of usable groundwater storage, TGS is computed for percentages ($p\%$) of V to determine the number of years until the volume in storage is depleted by p percent. We quantify the number of years until the study aquifers are depleted to thresholds set at 25% and 90% of total capacity. For example, $TGS_{90\%}$ computes the number of years until the total volume in storage is depleted by 90% of the total storage capacity. The percentage values can also account for the influence of depletion that may have occurred prior to the study period. Supporting information Text S2 provides additional discussion on the influence of prior depletion for the aquifers that have available long-term depletion estimates.

3. Results

3.1. Total Groundwater Storage

3.1.1. Comparison of Historical and Regional Estimates

Figure 3 summarizes the range of storage estimates representing the groundwater availability input in the Total Groundwater Stress (TGS) ratio for the depleting aquifers. For each aquifer, estimates delineate historical estimates (left), regional estimates from section 2.2.2 (middle) and revised estimates (right) based on ranges of specific yield and saturated thickness as described in section 2.2.3. For nine of the aquifers, the regional estimates and revised estimates are within a similar range. In comparison, the historical estimates are above the range of either the regional or revised estimates. Eight of these depleting aquifers do not have regional estimates for comparison. This suggests that the revised estimates can provide a new baseline storage estimate based on hydrogeologic parameters for the aquifer systems that lack regional estimates. The range of storage estimates is summarized in supporting information Table S1.

Figures 4a and 4b show the distributed historical estimates of storage in the study aquifers. Within the historical estimates, the differences between the lower and upper limits of storage vary by a factor between about

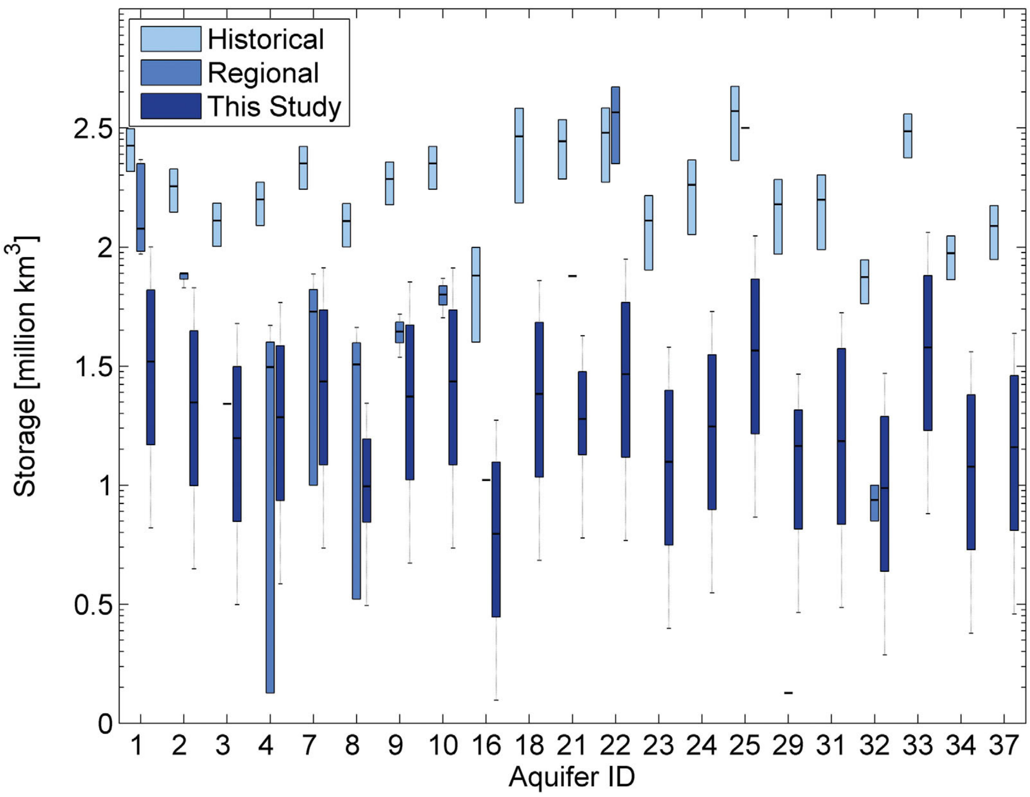

Figure 3. Estimates of total storage (million km^3) in the depleting study aquifers on a log scale based on the historical, regional, and revised estimates of storage ("This Study"). The median is present within the boxplots. The outliers have been removed. Aquifers with a single estimate of storage in a category are marked with a single median marker (–).

2 and 6. The discrepancy is a function of the saturated thickness value as either 1000 or 2000 m and the value of porosity assumed as a constant 1% or varying with depth. The difference between storage estimates is increased when comparing the historical and regional estimates in the 23 study aquifers with independent estimates of storage (Figure 4c). A comparison between Figures 4c and 4a–4b highlights the large discrepancy between the commonly cited historical estimates and the regional case studies, as they differ over 3–4 orders of magnitude (supporting information Table S1). For example, *Swezey* [1999] cite the volume of storage in the Northwest Sahara Aquifer System (Aquifer #2, "Sahara") as between 28 and 70 km^3 as compared to 450,000 km^3 from the distributed estimate by *Korzun* [1978]. This comparison shows that the majority of the historical estimates provide large overestimations of the volume of water in storage when compared to available regional estimates.

3.1.2. Revised Estimates

The revised estimates constrain the range of historical and regional estimates of aquifer storage. Figures 4d–4f select three combinations of equations (6) and (7) to illustrate the influence of changing hydrogeologic constraints from the historical methods. The remaining combinations of equations (6) and (7) are discussed below and listed in supporting information Table S1. Figure 4d shows storage in the study aquifers according to equation (6), where storage is the product of the minimum specific yield, 200 m saturated thickness, and the aquifer area. Figures 4e and 4f maintain the 1% porosity assumption from *Nace* [1969], with aquifer area and saturated thickness of 200 and 1000 m, respectively.

A comparison is made between the minimum regional estimates and the revised estimates to determine the combination of hydrogeologic characteristics that produce the closest estimate to regional values. Supporting information Table S1 shows which hydrogeologic characteristics can be combined to reproduce the regional storage estimates. We find that the majority of the study aquifers require the minimum specific yield estimate in equation (6) or a saturated thickness less than 500 m from equation (7). For example, the 110 km^3 storage capacity estimate in the Sudd Basin (Aquifer #8, "Sudd") [*Margat and van der Gun*, 2013] can be reproduced with an assumed 1% porosity and a saturated thickness between 20 and 50 m. This is 2 orders of magnitude

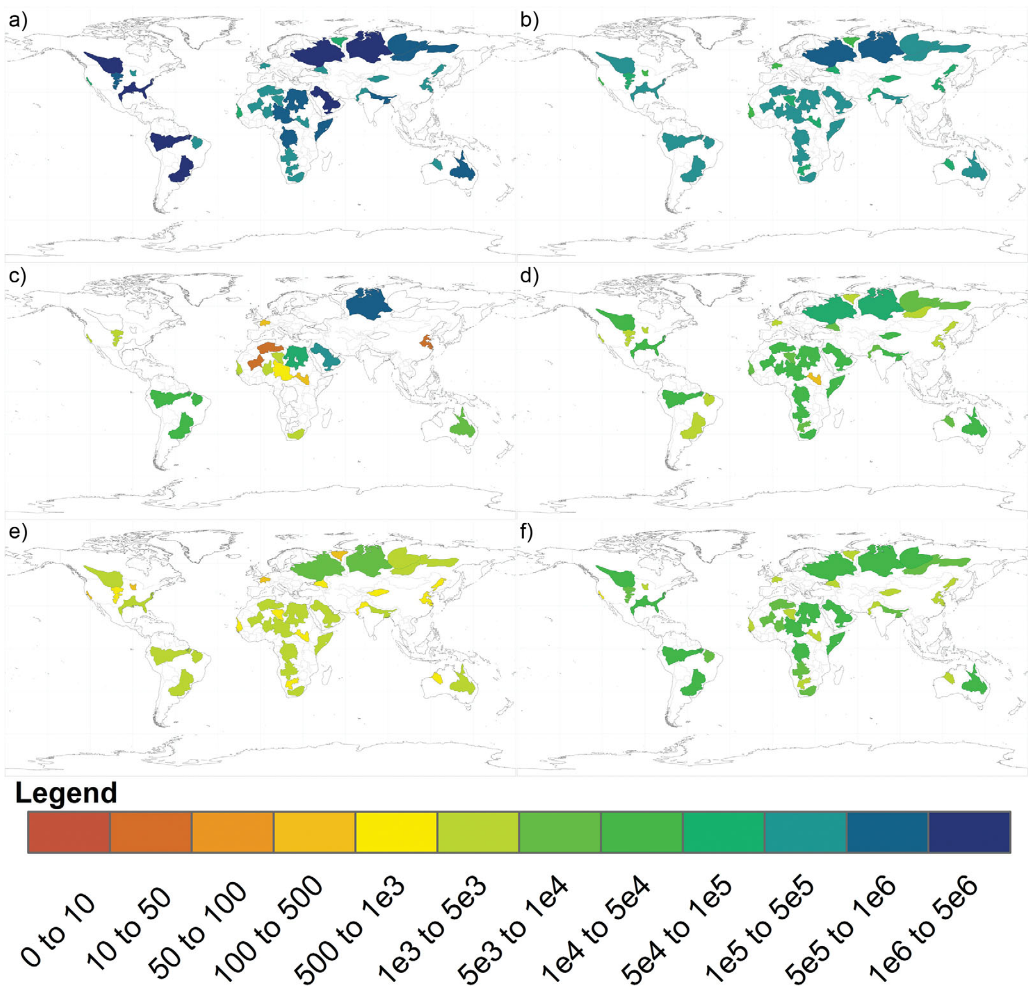

Figure 4. Estimates of total storage in each study aquifer based on the historical, regional, and revised estimates in cubic kilometers. (a) Distributed upper historical limit by *Korzun* [1978], (b) distributed lower historical limit by *Nace* [1969], (c) the minimum available regional estimate, (d) revised storage estimated according to equation (6) with minimum estimate of specific yield, (e) revised storage estimated according to equation (7) with 200 m saturated thickness, (f) revised storage estimated according to equation (7) with 1000 m saturated thickness. Outlined aquifers without colors indicate systems that lack available regional storage estimates. Note that, many cases, the historical estimates of *Nace* [1969] and *Korzun* [1978] are 1–3 orders of magnitude larger than the regional estimates, and as such, we place little confidence in them. See supporting information Table S1 and the text for more information.

less than the assumed saturated thickness by *Nace* [1969] and *Korzun* [1978]. Additionally, *Margat and van der Gun* [2013] report a storage range for the Paris Basin (Aquifer #32, "Paris") as between 500 and 1000 km^3. The reported range for the Paris can be reproduced with 1% porosity and between 200 and 500 m saturated thickness, less than half the thickness reported by *Nace* [1969] and *Korzun* [1978]. These discrepancies imply that these historical assumptions result in a severe overestimation of aquifer storage.

3.2. Distribution and Severity of Total Groundwater Stress

3.2.1. Total Groundwater Stress

Figures 5 and 6 present Total Groundwater Stress (TGS) as the ratio of total groundwater storage to groundwater depletion for 25% and 90% depletion, respectively. The TGS ratio results in the number of years to depletion, in this case assuming depletion continues at a constant rate into the future. Supporting information Tables S2 and S3 summarize the results of TGS for the range of storage estimates presented in this study. The TGS ratio varies over orders of magnitude within an aquifer system, depending on the storage estimate used, indicating that the uncertainty in aquifer storage severely limits the calculation of aquifer resilience.

Supporting information Table S4 shows that the depletion, storage, and TGS estimates from the current study are within the range of available published estimates for eight study aquifers. Supporting information

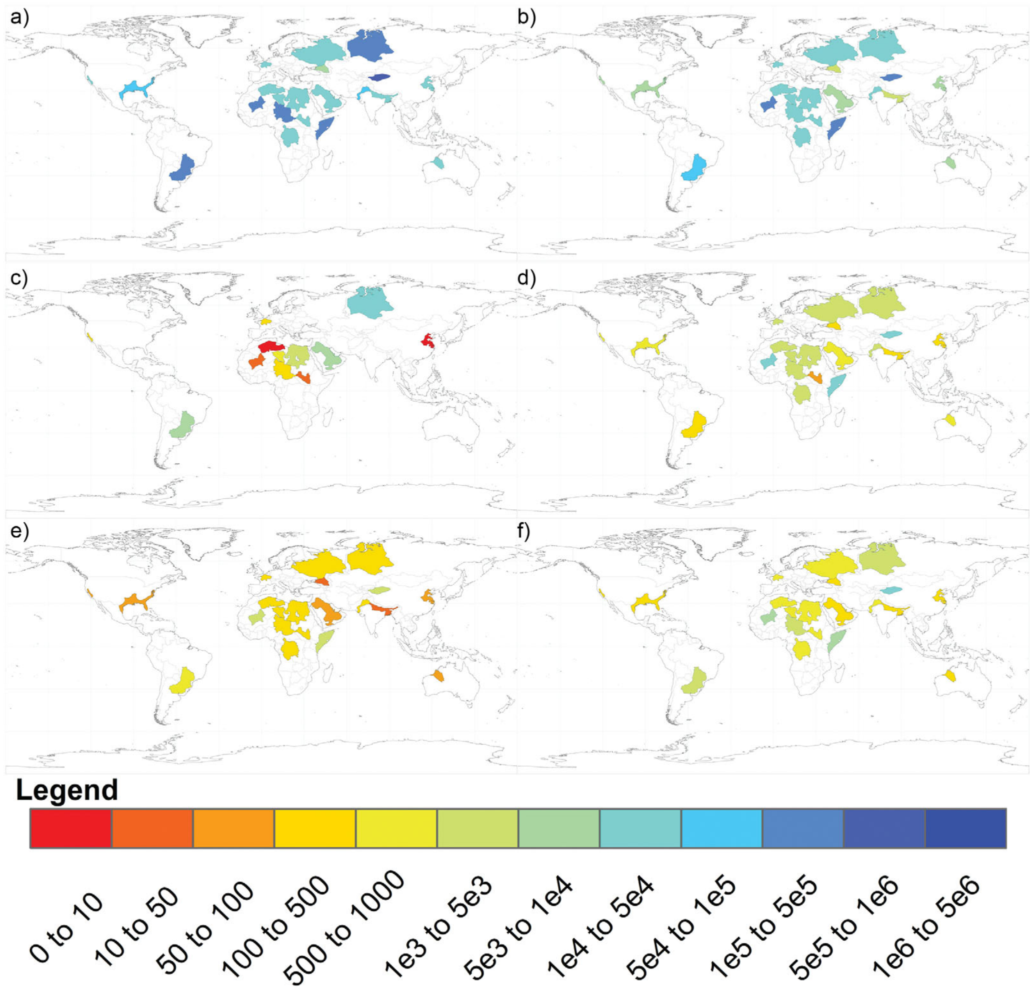

Figure 5. Total Groundwater Stress as the number of years to 25% depletion ($TGS_{25\%}$). (a) Distributed upper historical limit by *Korzun* [1978], (b) distributed lower historical limit by *Nace* [1969], (c) the minimum available regional estimate, (d) revised storage estimated according to equation (6) with minimum estimate of specific yield, (e) revised storage estimated according to equation (7) with 200 m saturated thickness, (f) revised storage estimated according to equation (7) with 1000 m saturated thickness. See text for discussion of lack of confidence in TGS estimates using the historical storages from *Nace* [1969] and *Korzun* [1978].

Text S2 presents a detailed discussion of this comparison and assesses the influence of depletion prior to the study period on the TGS estimates. Prior depletion was found to have the greatest influence in the Central Valley, resulting in a 12% decrease in TGS. Prior depletion had negligible influence on the remaining aquifers with records of 20th century depletion, discussed further in supporting information Text S2.

The Ganges has the highest rate of depletion from GRACE of 19.6 ± 1.2 millimeters per year (mm/yr) (12.2 ± 0.8 km^3/yr). The TGS ratio in the Ganges ranges from approximately 10 years to 90% depletion based on the lowest revised estimate of storage to nearly 10,000 years with the lower historical estimate. Conversely, the Tarim Basin (Aquifer #31, "Tarim") has the lowest depletion rate by GRACE of 0.23 ± 0.3 mm/yr (0.11 ± 0.1 km^3/yr) and low water availability based on recharge, the lower historical estimate, and the revised estimate by equation (7). The Tarim is a small aquifer by area, but the low depletion rate results in approximately 800 years to 90% depletion by the smallest storage estimate.

3.3. A Comparison of Total and Renewable Groundwater Stress

Renewable Groundwater Stress (RGS) as defined by *Richey et al.* [2015] evaluates aquifer stress resulting from renewable groundwater availability as the estimated groundwater recharge rate. Table 3 summarizes the differences in groundwater availability in the study aquifers depending on whether recharge or storage

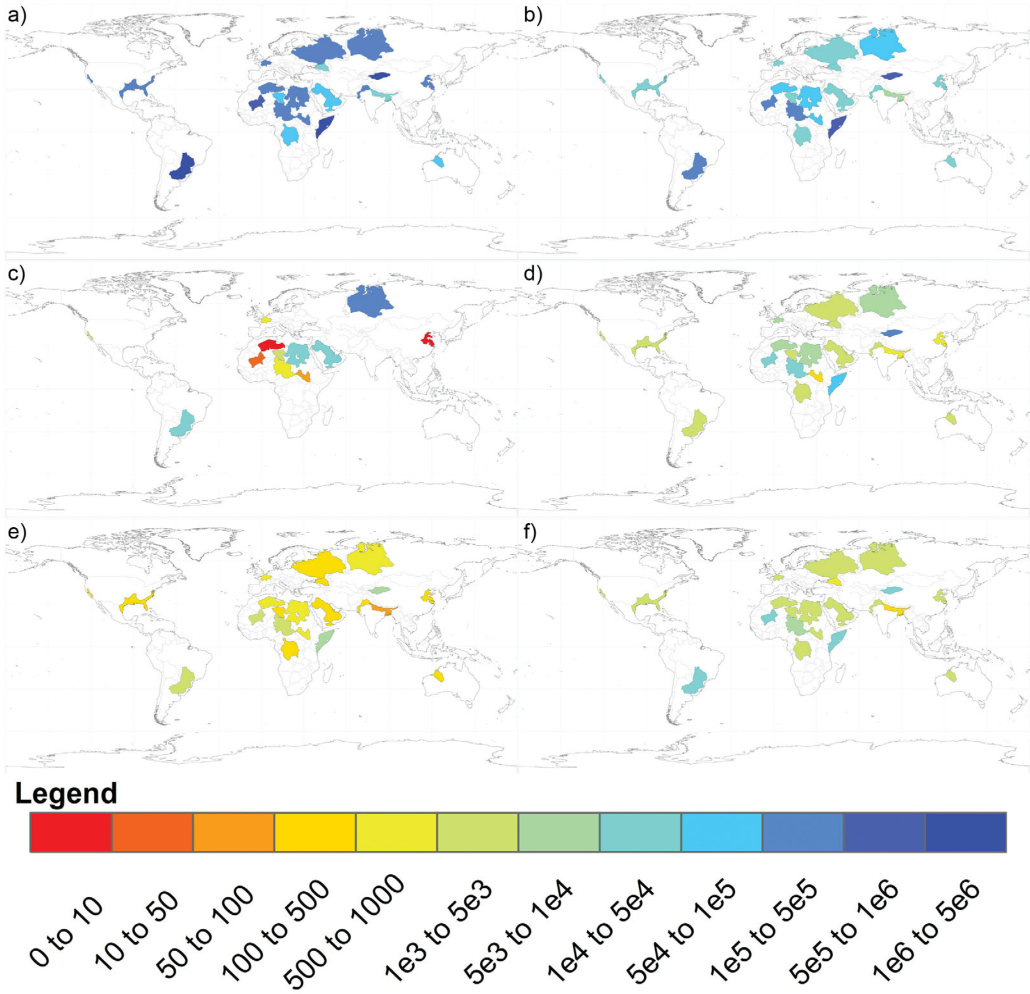

Figure 6. Total Groundwater Stress as the number of years to 90% depletion ($TGS_{90\%}$). (a) Distributed upper historical limit by *Korzun* [1978], (b) distributed lower historical limit by *Nace* [1969], (c) the minimum available regional estimate, (d) revised storage estimated according to equation (6) with minimum estimate of specific yield, (e) revised storage estimated according to equation (7) with 200 m saturated thickness, (f) revised storage estimated according to equation (7) with 1000 m saturated thickness. See text for discussion of lack of confidence in TGS estimates using the historical storages from *Nace* [1969] and *Korzun* [1978].

is used to define availability. It shows that the definition can change whether a system is considered to have limited or plentiful supplies. The Nubian Sandstone Aquifer System (Aquifer #1, "Nubian") has negligible renewable supplies but a large volume of water in storage.

The definition of water availability as storage or recharge further influences the assessment of groundwater resilience and stress, based on the RGS and TGS ratios. Although the estimate of groundwater depletion remains the same between the renewable [*Richey et al*., 2015, Figure 9] and total (Figures 5 and 6) stress ratios, the distribution and severity of each type of stress differs as a function of the definition of availability as a renewable flux or as a total storage volume. In our comparison, we only consider depleting aquifers to assess differences between RGS and TGS since aquifers that have positive trends in groundwater storage anomalies are considered to be resilient systems over the study period.

The aquifers that are considered overstressed or highly stressed from the RGS ratio and that remain highly stressed from the TGS ratio are indicative of aquifers that lack resilience due to high depletion with limited buffer capacity. There are eight aquifers that are overstressed by the RGS ratio, which is considered the least sustainable characteristic RGS stress regime. Only two of these systems remain highly stressed by TGS based on the minimum of regional estimates, including the Sahara. The Sahara has about 10 years to 90%

Table 3. A Comparison of Groundwater Availability Estimates Defined as Mean Annual Recharge and as Total Storage[a]

Aquifer ID	Mean Annual Recharge (mm)	Aquifer ID	*Nace* [1969] (km^3)	Aquifer ID	*Margat and van der Gun* [2013]-Minimum (km^3)	Aquifer ID	Equation (6): Minimum Sy, b = 200 m (km^3)	Aquifer ID	Equation (7): b = 20 m (km^3)	
8	−18.4	16	16,000	4	18	8	490	16	16	LOW AVAILABILITY
12	−12.3	15	28,000	29	18	15	1,100	15	27	
13	−11.8	32	33,000	8	110	16	1,600	32	38	
9	−5.9	35	40,000	32	500	21	1,800	35	44	
23	−4.6	5	41,000	7	1,000	30	2,100	30	53	
17	−3.7	34	53,000	16	1,100	29	3,400	5	55	
22	−2.6	30	54,000			32	3,800	34	57	
31	−0.7	23	64,000			17	4,100	23	63	
1	−0.3	37	79,000			35	4,400	37	83	
2	−0.3	29	87,000			20	4,500	29	85	
3	−0.2	12	88,000			27	4,800	12	86	
4	1.0	31	95,000			5	5,500	31	93	
7	6.0	3	100,000	5	1,500	34	5,700	8	98	
37	6.1	8	100,000	13	3,000	23	6,300	3	99	
14	8.4	17	100,000	3	4,800	28	6,900	17	100	
6	11.2	20	110,000	36	8,700	37	8,300	20	110	
36	13.7	6	120,000	6	10,000	12	8,600	6	120	
28	16.5	13	120,000	17	15,000	3	9,900	13	120	
10	19.0	27	120,000	20	18,000	26	9,900	24	120	
30	20.2	24	130,000			6	12,000	27	120	
16	24.1	4	150,000			13	12,000	4	150	
34	28.8	28	170,000			24	12,000	28	170	
5	34.4	2	200,000			31	14,000	2	200	
26	36.2	11	210,000			36	14,000	11	200	
27	36.4	9	220,000			4	15,000	9	220	
25	39.4	18	230,000			2	20,000	18	230	HIGH
29	96.6	26	250,000			11	20,000	26	250	
33	98.6	7	300,000			9	22,000	7	300	
11	101.1	10	300,000			18	23,000	10	300	
32	133.6	14	310,000			7	30,000	14	300	
15	151.8	22	350,000			10	30,000	22	340	
35	161.3	36	350,000	19	32,000	14	30,000	36	350	
18	168.4	21	370,000	21	57,000	22	34,000	21	360	
24	214.4	1	430,000	2	60,000	1	44,000	1	440	
21	225.7	19	470,000	1	540,000	19	46,000	19	460	
20	323.0	25	530,000	25	1,000,000	25	54,000	25	540	
19	546.6	33	560,000	22	2,200,000	33	57,000	33	570	

[a]The determination of low (high) availability is determined as the lowest (highest) third of available supplies across the study aquifers. A negative value of recharge refers to systems that are dominated by the upward flow of capillary fluxes away from the water table as opposed to the positive (downward) flux of recharge.

depletion from the minimum regional estimate of storage. However, the uncertainty in storage estimates is highlighted in the Sahara where $TGS_{90\%}$ ranges from about 10 years to 150,000 years within different regional estimates.

The majority of the aquifers characterized by low renewable stress have high levels of total stress. The Ganges is a highly depleting system, but has a high rate of mean annual recharge that result in low renewable stress ratios. These systems have low numbers of years to 90% depletion based on the revised estimates of storage, but there is no regional estimate for comparison. These aquifers are vulnerable to increases in depletion or decreases in recharge that might result in a shift from low to high renewable stress conditions, which could pressure the long-term buffer capacity of the aquifer.

4. Discussion

We show that a wide range of variability exists in estimates of total storage, leading to great uncertainty in the state of global groundwater stocks. Even within a single aquifer, storage estimates vary over multiple orders of magnitude. This finding supports the warnings by *Shiklomanov* [1993] and *Famiglietti* [2014] that the historical estimates providing our current knowledge of global groundwater storage are inaccurate. The uncertainty range clearly indicates that in most cases, we do not know how much groundwater exists in storage to maintain unsustainable groundwater depletion. Therefore, the ability to quantify aquifer resilience is severely limited.

The previous state of knowledge on groundwater stocks relied on the historical estimates that likely overestimate groundwater volume for the study aquifers by multiple orders of magnitude. The historical estimates, based on the assumption that there is a constant and extensive groundwater supply across the global land surface, create a severe misrepresentation of the volume of global groundwater. Such overestimates can lead to an assumption that groundwater is an infinite resource: as such they should no longer be blindly accepted as realistic estimates of groundwater storage. By comparing historical and regional estimates of storage, we show that such an assumption is not valid in all of the study aquifers, for example, in the Sahara, Taoudeni-Tanezrouft Basin (Aquifer #4, "Taoudeni"), and North China Aquifer System (Aquifer #29, "North China") with minimum reported regional estimates of less than 30 km^3. Simple hydrogeologic assumptions were made that constrain the estimates of groundwater storage to within a range that more closely matches available regional estimates. Aquifer-specific estimates of saturated thickness and specific yield based on *in situ* observations are essential to further constrain the revised storage estimates and provide realistic TGS ratios.

The influence of total storage on aquifer resilience is highlighted by a comparison between the Nubian and the Sahara. The depletion rates in the aquifers are similar at 6.08 $\pm$ 1.9 and 2.69 $\pm$ 0.8 cubic kilometers per year (km^3/yr), respectively. The depletion rate in the Sahara is comparable to the reported rate of greater than 2.2 km^3/yr by *Mamou et al*. [2006]. The GRACE-based depletion rate in the Nubian is greater than the value of 2.17 km^3/yr from *Bakhbakhi* [2006], although this value was reported for the year 2000 and has not been updated for the GRACE period. The overstressed Renewable Groundwater Stress (RGS) ratios of 10.6 for the Nubian and 10.8 for the Sahara are also similar between the aquifers. However, the estimates of total storage in each aquifer result in contrasting estimates of TGS. The minimum regional estimate of storage in the Sahara is 28 km^3, which results in about 10 years to 90% depletion. An independent storage estimate from *Mamou et al*. [2006] is not available as a comparison point to our estimate of depletion timescales. Conversely, the minimum regional estimate of storage in the Nubian is 150,000 km^3, which results in a buffer capacity that is 4 orders of magnitude greater than in the Sahara. *Bakhbakhi* [2006] shows the usable lifespan of the Nubian diminishes exponentially as a function of current extraction rates and the remaining volume of water in storage that the author considers recoverable, limited by rising extraction costs, and decreasing water quality. Such a comparison suggests the importance of accounting for both renewable fluxes and total stocks when assessing the sustainability of groundwater use in an aquifer system.

The Ganges and North China systems represent an alternative case. Despite high rates of depletion, high rates of mean annual recharge place both aquifers in the low stress category of the Variable Stress RGS regime. However, low regional and revised storage estimates result in a limited number of years to 90% depletion. The depletion estimate in the North China system by *Liu et al*. [2011] of 3.52 km^3/yr is within the GRACE-derived depletion error range of 3.2 $\pm$ 0.6 km^3/yr. The storage estimate by *Liu et al*. [2011] is also comparable to the minimum estimate used in our study, at 23.8 km^3. These are systems that have limited resilience and buffer capacity that could make them vulnerable to increased rates of depletion and decreases in renewable available supplies. Additionally, the three aquifers with the highest rates of depletion do not have regional estimates of storage to provide a further constraint on available supplies.

The timescales presented here offer maximum estimates of TGS. The study period is limited by the length of the GRACE satellite record. In the results presented here, we implicitly assume that the volume of storage in the study aquifers is at full capacity at the start of the study period in 2003. In reality, pumping has already been occurring, leaving legacy effects on the system [*Liu et al*., 2007] such that the years to depletion are less than indicated by our results (supporting information Text S1). We assume the range of percentage values (*p%*) of storage provides baseline depletion timescales that encompass the influence of decreasing accessibility and usability of storage with continued external pressures that, as well as the legacy effects of depletion prior to the study period, could alter TGS timescales. It is possible that the 25% depletion estimates still provide an optimistic estimate of depletion timescales, for example, in the Nubian where the volume of recoverable freshwater in storage is less than 3% of the total storage capacity [*Bakhbakhi*, 2006]. Accounting for the difference in recoverable freshwater versus total storage reduces the time to depletion by a factor of 36 using the storage and depletion estimates of *Bakhbakhi* [2006] (supporting information Table S4). A further limit on available storage may exist due to regulatory structures, often associated with maintaining base flow to streams, such as in the Central Valley and High Plains [*Scanlon et al*., 2012].

It is critical to note that the spatial scale used in this study averages the changes in storage across the entirety of each study aquifer. Local and regional variations within the aquifers can present a contrasting picture of total stress and time to depletion at a smaller scale or across national boundaries within an aquifer [*Wada and Heinrich*, 2013]. In the High Plains, for example, *Scanlon et al*. [2012] and *Famiglietti and Rodell* [2013] show that the northern High Plains is dominated by recharge while the southern High Plains is heavily pumped and parts of it could be depleted within 30 years. However, the spatial averaging across the aquifer in this study finds a near zero trend in groundwater storage change across the aquifer. In fact, many of the aquifers in this study have regions with more severe depletion rates that are balanced by less severe depletion or gaining storage. Previous GRACE studies that focus explicitly on high depletion regions instead of aquifer averages have higher magnitudes of depletion than occur in this study [e.g., *Rodell et al*., 2009; *Voss et al*., 2013].

We assume the rate of depletion will remain constant to provide a baseline estimate of TGS timescales, following the steady rise in groundwater depletion in large aquifers by *Konikow* [2011]. However, the rate will likely vary geographically as a function of socio-economic and physical factors [*Hardin*, 1968; *Dietz et al*., 2003; *Alcamo et al*., 2007]. For example, *Wada et al*. [2014] found a global average increase in groundwater use of about 3%/yr from 1990 to 2010, especially in North America, Central America, and parts of Asia, due largely to growing population and associated food demand. The combination of population growth and increased food demand with the potential for increased hydrologic extremes due to climate change may further increase the rate of use in groundwater systems as surface supplies become less accessible [*Kundzewicz and Döll*, 2009; *Famiglietti*, 2014]. These combined influences would act to shorten the timescales of depletion found in this study. Additionally, some regions, such as Sub-Saharan Africa have yet to experience an agricultural boom. Only about 5% of land is currently irrigated in this area, as opposed to the greater than 60% of irrigated land during India's Green Revolution [*Rockström et al*., 2007]. The expansion of irrigated agriculture has resulted in severe groundwater depletion in parts of India [*Rodell et al*., 2009] as well as an influx of arsenic in agriculture through increased irrigation of contaminated groundwater [*Brammer and Ravenscroft*, 2009]. Growing food demand and agricultural pressure may expand the need for irrigation and further increase the rate of depletion. Conversely, water-use efficiency practices could decrease the rate of depletion. While we recognize that increasing or decreasing the rate of future depletion can account for changes in climate and use patterns, it is beyond the scope of the present study.

5. Conclusions

The results presented herein explore the concepts of groundwater resilience and the buffer capacity of groundwater storage in a water stress framework. We define a Total Groundwater Stress (TGS) ratio as a measure of groundwater resilience in the world's largest aquifer systems. In this study, we highlight the state of knowledge on both fluxes and stocks in the world's largest aquifer systems at the aquifer scale. We compare available estimates of total storage, and further constrain these estimates, to assess groundwater stocks. Remote sensing observations from GRACE assess the trend in combined fluxes within a system, by integrating the influence of recharge, discharge, pumping, and capture into a single trend of water mass anomalies.

GRACE observations allow for the first-ever quantification of groundwater resilience by identifying the systems that can no longer increase capture to balance external perturbations to an equilibrium state. The GRACE-based estimates of depletion integrate the dynamic, nonlinear links that exist in coupled human-natural systems like groundwater. This is necessary in a groundwater sustainability study to account for both human actions such as pumping and the dynamic response of the aquifer [*Liu et al*., 2007; *Zhou*, 2009]. Traditionally, the study of such a coupled system has been limited to either the human or natural dimension, though fully assessing resilience must account for both dimensions [*Liu et al*., 2007]. The spatial scale of GRACE allows for system-wide basin averages to address the resilience across the totality of the system as recommended by *Turner et al*. [2003].

Long-term storage loss is the limit of groundwater resilience, indicating an aquifer system's inability to maintain equilibrium despite perturbations. Groundwater is largely unregulated, although the influence of human management is often required to improve the resilience of natural systems [*Liu et al*., 2007]. An aquifer that increases capture by decreasing base flow may not be considered resilient in a coupled surface water-groundwater system, increasing the importance of management to improve system-wide resilience. Transparent information exchange on the state of fluxes and stocks in common pool resources, such as groundwater, is the first step toward effective management [*Schlager et al*., 1994; *Dietz et al*., 2003]. We have shown that transparent knowledge on groundwater stocks is lacking in the majority of the world's large aquifer systems.

This work clearly demonstrates that it is no longer adequate to continue citing decades-old, heuristically derived, highly uncertain estimates of total groundwater storage. The lack of ground-based measures of total storage will continue to prevent a full characterization of aquifer stress and resilience until large scale efforts are implemented to improve the state-of-knowledge on groundwater stocks. To improve current storage estimates require a significant investment in regional monitoring and measuring systems to better characterize saturated thickness and soil properties within an aquifer. *Konikow* [2011, 2013] cites the most reliable methods to assess the state of a groundwater system as using observations of groundwater levels and storage coefficients, with temporally varying observations of gravity, and with a calibrated model. *Famiglietti* [2014] calls for detailed hydrogeologic exploration of the world's major aquifers. All of these methods require *in situ* observations for direct measurements, model calibration, and an assessment of subregional conditions. Without these measurements, the extrapolations of limited data across large land areas is required often with a high level of uncertainty. The continuation of remote sensing missions is crucial to provide an integrated perspective of the combined human and natural influences on a groundwater system [*Famiglietti and Rodell*, 2013]. Improved assessments of soil moisture over large scales [*Entekhabi et al*., 2010] would benefit the isolation of groundwater storage changes from estimates of total terrestrial water storage. Additionally, it is important to incorporate decision makers into an assessment of recoverable storage capacity for transparency on the quality of information regarding available supplies and to ultimately create water use regulations that holistically and sustainably address the combined human and natural impacts on a groundwater system.

Until improved storage estimates exist to determine a system's full capacity to buffer against Renewable Groundwater Stress, continued pressure on aquifer systems could lead to irreversible depletion that seriously threaten the sustainability of groundwater dependent regions. Additionally, large volumes of water in storage may not be representative of resilient systems without also considering the negative impacts of pumping and limitations on recoverable storage as a function of soil properties and well design, which reduce the usable storage volume.

The uncertainty in total groundwater storage and estimates of depletion timescales is particularly relevant in regions that are prone to drought and lack active management of groundwater resources. For example, the highly stressed Central Valley lacks sufficient natural recharge to balance current use rates [*Richey et al*., 2015], which is exacerbated by an increased dependence on groundwater during drought [*Famiglietti et al*., 2011; *Scanlon et al*., 2012]. The first regulations to govern groundwater use across the state were passed in 2014 and do not require sustainable groundwater management until 2040. The current depletion rate shows that the aquifer is unable to balance the combined impact of groundwater use and drought, either through capture or active management, and is therefore not a resilient system. The best available estimates of total storage in the Central Valley trace to estimates made in the 1970s and 1980s [*Department of Water Resources (DWR)*, 1975; *Williamson et al*., 1989]. The earlier of these studies estimated the volume of recoverable storage at that time to be 176 km^3 versus 1600 km^3 of total capacity [*DWR*, 1975], implying the lifespan of usable groundwater today that is highly threatened and may be expended in a matter of decades given current rates of groundwater depletion [*Famiglietti et al*., 2011]. This work highlights the need to improve active management of groundwater by both reducing demand and by increasing supply, for example, through artificial recharge [*Scanlon et al*., 2012].

Here, we highlight the ability to provide bounds on groundwater resilience and buffer capacity based on available and constrained storage estimates in the study aquifers. We show how remote sensing observations from GRACE can improve our understanding of groundwater stress and resilience by quantifying depletion; however, the large uncertainty in storage remains a barrier. As continued efforts increase the transparency and availability of information on the state of large aquifer systems, the ability to manage these systems to increase resilience will be enhanced.

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