WATER RESOURCES RESEARCH, VOL. 45, W04408, doi:10.1029/2008WR007333, 2009

Basin-scale, integrated observations of the early 21st century multiyear drought in southeast Australia

Marc J. Leblanc,[1] Paul Tregoning,[2] Guillaume Ramillien,[3] Sarah O. Tweed,[1] and Adam Fakes[4]

Received 4 August 2008; revised 10 November 2008; accepted 26 December 2008; published 8 April 2009.

[1] The Murray-Darling Basin in southeast Australia is experiencing one of the most severe droughts observed recently in the world, driven by several years of rainfall deficits and record high temperatures. This paper provides new basin-scale observations of the multiyear drought, integrated to a degree rarely achieved on such a large scale, to assess the response of water resources and the severity of the drought. A combination of Gravity Recovery and Climate Experiment (GRACE) data with in situ and modeled hydrological data shows the propagation of the water deficit through the hydrological cycle and the rise of different types of drought. Our observations show the rapid drying of soil moisture and surface water storages, which reached near-stationary low levels only ~2 years after the onset of the drought in 2001, with a loss of ~80 and ~12 km^3 between January 2001 and January 2003, respectively. The multiyear drought has led to the almost complete drying of surface water resources which account for most of the water used for irrigation and domestic purposes. High correlation between observed groundwater variations and GRACE data substantiates the persistent reduction in groundwater storage, with groundwater levels still declining 6 years after the onset of the drought (groundwater loss of ~104 km^3 between 2001 and 2007). The hydrological drought continues even though the region returned to average annual rainfall during 2007.

Citation: Leblanc, M. J., P. Tregoning, G. Ramillien, S. O. Tweed, and A. Fakes (2009), Basin-scale, integrated observations of the early 21st century multiyear drought in southeast Australia, *Water Resour. Res.*, *45*, W04408, doi:10.1029/2008WR007333.

1. Introduction

[2] Droughts affect more people globally than any other natural hazard [*Bryant*, 2005]. A drought is regional by nature and it is characterized by its water deficit, yet quantifying total water deficit over large areas is a major challenge in drought studies. We present observations at the basin scale of the total water deficit (including surface water, soil moisture and groundwater) in the ~1 million km^2 Murray-Darling Basin in southeast Australia during the ongoing multiyear drought that commenced in 2001.

[3] The Murray-Darling drainage basin accounts for approximately 20–30% of the gross value of Australia's agricultural production [*Ritchie et al.*, 2004; *Van Dijk et al.*, 2007] and includes ~60% of Australia's irrigated land [*Van Dijk et al.*, 2007]. Nearly 67% of the basin is agricultural land which is used for pasture and cropping and 32% is native forest (Australian Bureau of Rural Sciences data; available at http://adl.brs.gov.au/water2010). From 2001 to 2006, the total rainfall deficit in the basin compared to the long-term (1900–2006) mean annual rainfall is estimated at ~520 km^3 (Figure 1) and maximum rainfall deviations (~190 mm) from the long-term annual mean occurred in 2002 and 2006 (Australian Bureau of Meteorology data; available at http://www.bom.gov.au). In September 2007, the total volume of surface water stored in the River Murray system in the southern part of the basin was only ~2 km^3, or 23% of the total capacity. The ongoing drought has resulted in low or zero irrigation allocations, causing significant impacts on irrigators and their communities. A loss in biodiversity is recorded with many water bodies drying out and water quality degrading.

[4] There is a need for basin-wide observation and integrated water accounting in order to understand the severity of the current water crisis in southeast Australia. Droughts are typically characterized into categories including "meteorological," "hydrological," "agricultural" and "socioeconomic" [*Wilhite and Glantz*, 1985; *Keyantash and Dracup*, 2002], yet it remains difficult to measure integrated bulk water deficits over large regions [*Heim*, 2000; *Wilhite*, 2000; *Keyantash and Dracup*, 2002]. Water can be stored in basins as surface water, soil moisture, groundwater, snow/ice and in vegetation. Most of these forms are difficult, if not impossible, to measure accurately at any macroscale; therefore, drought indicators rely upon proxies and approximations to quantify the integrated

[1]Hydrological Sciences Research Unit, School of Earth and Environmental Sciences, James Cook University, Cairns, Queensland, Australia.
[2]Research School of Earth Sciences, Australian National University, Canberra, ACT, Australia.
[3]Dynamique Terrestre et Planétaire, Observatoire Midi-Pyrénées, UMR5562, CNRS, Toulouse, France.
[4]CSIRO Sustainable Ecosystems, James Cook University, Cairns, Queensland, Australia.

0043-1397/09/2008WR007333$09.00

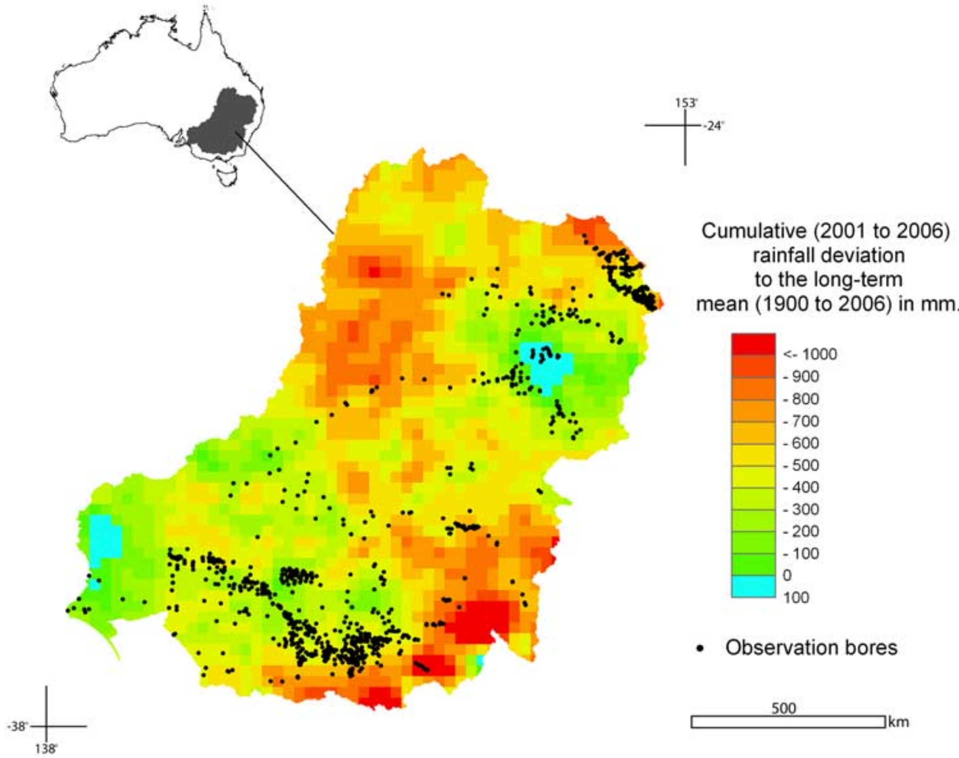

Figure 1. Cumulative rainfall deficit across the Murray-Darling Basin for the 2001–2006 period and location of the shallow groundwater monitoring bores.

temporal variations of water resources at basin scales [*Heim*, 2000; *Wilhite*, 2000; *Keyantash and Dracup*, 2004]. It is also problematic to characterize prolonged drought using proxies of rainfall and observations of surface water (flow rates, storage levels) because ongoing loss of deeper water resources can occur even after surface rivers and reservoirs have become dry. The only way to properly assess the impact of a drought on water resources is through an integrated measure of all water storage types. It is also important to assess and manage the response of water resources to drought at the basin scale, since the drought phenomenon is regional and a water deficit in a subcatchment can be propagated downstream. Until recently, water in the Murray-Darling Basin was managed by five states with a certain degree of independence. This created competition for limited water resources and imbalances in water allocations/diversions between upstream and downstream parts of the basin. Recent efforts by the Australian federal government are aimed at managing the water resources at the basin scale.

[5] The ability of the Gravity Recovery and Climate Experiment (GRACE) space gravity mission to monitor total terrestrial water storage with sufficient accuracy was demonstrated in a prelaunch assessment [*Rodell and Famiglietti*, 1999] and subsequently in different basins across the world [e.g., *Ramillien et al.*, 2004; *Wahr et al.*, 2004]. An analysis of ground-based measurement collected from 2002 to 2003 in a large catchment (Murrumbidgee) in the southwest of the Murray-Darling Basin indicates that changes in water storage are within a range for which GRACE should be able to provide a statistically significant measure [*Ellett et al.*, 2006]. GRACE data have been used to quantify evapotranspiration in the Mississippi Basin [*Rodell et al.*, 2004a] and to estimate continental water variability at seasonal scale [*Ramillien et al.*, 2005; *Schmidt et al.*, 2006]. In these studies, time series and maps of water storage estimates were computed from GRACE data as regional averages over areas of at least $\sim$1 million km^2. GRACE-based time series of regional water storage have been compared in different large drainage basins and generally agree well with outputs from global hydrological models, especially at an annual time scale [*Schmidt et al.*, 2006]. GRACE has been used to detect drying trends in large basins [*Crowley et al.*, 2006; *Syed et al.*, 2008] and has been used to quantify the effect of the 2003 heat wave in central Europe [*Andersen et al.*, 2005] on continental water storage.

[6] This paper aims to present new observations of a multiyear drought, integrated to a degree that has not been achieved previously on such a large scale. Globally, observations of the impact of multiyear droughts on water resources are scarce, even more so at basin scale and for semiarid areas which often have limited in situ monitoring networks [*Wilhite*, 2000; *Tallaksen and Van Lanen*, 2004]. We used a combination of GRACE observations with in situ and modeled hydrological data to characterize the multiyear drought in the Murray-Darling Basin and to monitor its relative impact on the major water stores at basin scale. We computed from GRACE data the total water deficit and quantified the severity of the multiyear drought. Additionally, to monitor the propagation of the drought in the terrestrial branch of the hydrological cycle we used (1) in situ monitoring networks to estimate monthly and annual change in surface water and groundwater storages and (2) outputs of a land surface model to estimate monthly soil moisture storage.

Results are discussed in terms of implications for improving our understanding of water resources response to natural climate variability and anthropogenic change.

2. Methodology

[7] The temporal change in terrestrial water storage, ΔTWS, stored in a basin can be written as the sum of the different reservoir contributors:

$$\Delta TWS = \Delta SW + \Delta SM + \Delta GW \tag{1}$$

where SW represents the total surface water storage including lakes, reservoirs, farm dams and in-channel water (snow layer not included in Murray-Darling Basin); SM is the total soil moisture in the storage depth of the soil, GW is the total groundwater storage in the aquifers. These terms are generally expressed in volume (km^3) or mm of equivalent water height.

2.1. GRACE

[8] The GRACE satellites detect temporal changes in the Earth's gravity field. These changes are related to geophysical processes such as glacial isostatic adjustment but also to the redistribution of surface loads such as continental water storage. We used GRACE observations to estimate variations in TWS over the Murray-Darling Basin from 2002.6 to 2008.4 using a spherical harmonic representation of global-scale changes in the Earth's gravity field. We used the monthly geoid coefficient solutions (to degree 50 which equates to ~400 km on the surface of the Earth) of the Groupe de Recherche en Géodesie Spatiale (GRGS) version release 1. These solutions are generated with an epoch spacing of 10 days using a span of 30 days of observations [*Lemoine et al.*, 2007]. The effects of atmospheric mass, ocean tides and barotropic signals are accounted for using European Centre for Meteorological Weather Forecasting reanalysis, Finite Element Solution 2004 (FES2004) [*Le Provost et al.*, 1998] and the MOG2D-G barotropic [*Carrère and Lyard*, 2003] models, respectively. The remaining signals should correspond mainly to continental water storage changes over regions such as Australia where there are no significant tectonic or glacial isostatic adjustment signals. Geoid coefficients were obtained by multiplying the dimensionless Stokes coefficients by R, the Earth's mean radius (6371 km). We removed the mean gravity field from the spherical harmonic coefficients at each 10-day epoch and converted the geoid anomaly coefficients δU_{nm} and δV_{nm} into equivalent water height coefficients δC_{nm} and δS_{nm} using the following relation [e.g., *Ramillien et al.*, 2006]:

$$\begin{pmatrix} \delta C_{nm}(\Delta t) \\ \delta S_{nm}(\Delta t) \end{pmatrix} = \left[\frac{4\pi G \rho_w R}{(2n+1)\gamma}\left(1+k'_n\right)\right]^{-1} \times \begin{pmatrix} \delta U_{nm}(\Delta t) \\ \delta V_{nm}(\Delta t) \end{pmatrix} \tag{2}$$

where k'_n are the elastic Love load numbers, and n and m are the harmonic degree and order of the spherical harmonic fields, G is the gravitational constant (6.673 × 10^{-11} m^3 kg^{-1} s^{-2}), ρ_w is the mean density of water (~1000 kg m^{-3}) and γ is the mean gravity acceleration (9.81 m s^{-2}).

[9] We computed the corresponding variation of water volume of the basin, $\delta\Psi$ (Δt), as the scalar product of the water mass coefficients δC_{nm}, δS_{nm} with A_{nm} and B_{nm} the normalized harmonic coefficients of the Murray-Darling Basin geographical mask [e.g., *Ramillien et al.*, 2006]:

$$\delta\Psi(\Delta t) = 4\pi R^2 \sum_{n=1}^{N} \sum_{m=0}^{n} \{A_{nm}\delta C_{nm}(\Delta t) + B_{nm}\delta S_{nm}(\Delta t)\} \tag{3}$$

[10] In the following, GRACE-based TWS variations are expressed in terms of water volume changes $\delta\Psi$, or equivalent water height changes if $\delta\Psi$ is divided by the area of the drainage basin. We can now calculate ΔTWS as the variation of the water mass inside the drainage basin area between different epochs, t_1 and t_2:

$$\Delta TWS = TWS(t_2) - TWS(t_1) \tag{4}$$

2.2. Surface Water

[11] In the predominantly semiarid Murray-Darling Basin, most of the surface water is regulated using a network of reservoirs, lakes and weirs [*Kirby et al.*, 2006] and the surface water stored in these systems represent most of the total surface water (SW) present across the basin. A daily time series of the total surface water storage in the network of reservoirs, lakes, weirs and in-channel storage was obtained from the Murray-Darling Basin Commission and the state governments from January 2000 to April 2008 (Figure 2).

2.3. Groundwater

[12] Variations in groundwater storage can be estimated from in situ measurements as

$$\Delta GW = S_y \Delta \mathrm{H} \tag{5}$$

where S_y is the specific yield (ratio expressed as a percentage of the volume of water to the total volume of a rock that the rock will yield by gravity after being saturated) and H is the groundwater level (in meters) observed in monitoring bores. Estimates of annual changes in the total groundwater storage (GW) across the drainage basin were obtained from an analysis of groundwater levels in shallow monitoring bores from 2000 to 2007. The Murray-Darling drainage basin comprises several aquifers that can be regrouped into four categories: the Great Artesian Basin, the Murray Basin, the Darling regional groundwater basin and the fractured rock aquifers [*Ife and Skelt*, 2004]. Assuming that (1) the shallow aquifers across the Murray-Darling drainage basin are hydraulically connected and that (2) at a large scale the fractured aquifers can be assimilated to a porous media, changes in groundwater storage across the area can be estimated from observations of groundwater levels [*Rodell et al.*, 2007; *Strassberg et al.*, 2007]. Groundwater data were sourced from government departments of the states covered by the Murray-Darling Basin (QLD, Natural Resources and Mines; NSW, Department of Water and Energy; VIC, Department of Sustainability and Environment; and SA, Department of Water Land and Biodiversity Conservation). A total of 1462 representative bores for the various unconfined aquifers across the Murray-Darling Basin were selected on the basis of construction and monitoring details obtained from the state departments. Only observation bores (production bores excluded) with an average saturated zone ≤10 m from the

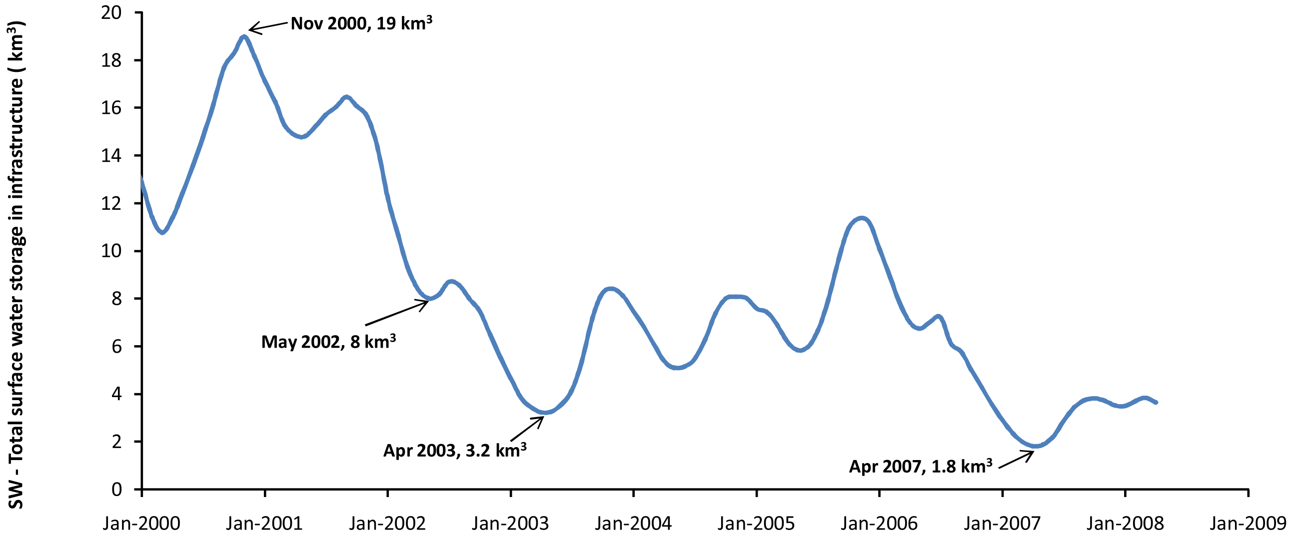

Figure 2. Change in surface water storage in the infrastructure system (reservoirs, lakes, weirs, and in-channel storage) of the Murray-Darling Basin.

bottom of the screened interval were selected. Deeper bores were excluded as they can reflect processes occurring on longer time scales [*Fetter*, 2001]. The potential influence of irrigation on some of the groundwater data is limited because during this period of drought irrigation is substantially reduced across the basin. 95% (1392) of the selected monitoring bores have a standard deviation of the seasonal groundwater levels below 2 m and were used to analyze the annual changes in groundwater storage during the period 2002 to 2007 (Figure 1). The remaining 5% of the bores, with the highest standard deviation, were discarded to exclude observation bores possibly under the immediate influence of local pumping or irrigation. Groundwater changes in the deep, confined aquifers (mostly GAB and Renmark aquifers) are due to either (1) a change in groundwater recharge (already accounted in water levels at their unconfined outcrop) or (2) deep pumping for farming (irrigation and cattle industry). Changes in *TWS* include those due to all water pumping, while changes in *GW* include those due to pumping in unconfined aquifers. Overall groundwater pumping across the basin was $\sim$1.6 km^3 in 2002–2003 [*Kirby et al.*, 2006] while pumping from the deep, confined aquifers was estimated to amount to 0.42 km^3 a^{-1} in 2000 [*Ife and Skelt*, 2004]. To allow direct comparison between *TWS* and *GW* estimates, pumping from the deep aquifers can be removed from the GRACE *TWS* time series assuming the 0.42 km^3 a^{-1} pumping rate remains constant during the study period.

[13] Changes in groundwater storage across the basin were estimated using an annual time step as most monitoring bores have limited groundwater level measurements in any year ($\sim$50% of bores with $\leq$4 measures per year). The annual median of the groundwater level was first calculated for each bore and change at a bore was computed as the difference of annual median groundwater level between two consecutive years. For each year a spatial interpolation of the groundwater level change was performed across the basin using a kriging technique. Spatially averaged groundwater level change for each year was reported as the deviation from the mean for the whole study period. An estimate of the total change in groundwater volume was derived using equation (5). The mean specific yield is estimated to be 0.05 for the shallow fractured rock aquifers [*Petheram et al.*, 2003], while it is estimated to range from 0.07 (sandy clay) to 0.18 (silts) for sedimentary aquifers [*Johnson*, 1967]. Therefore, values for the spatially averaged specific yield range between 0.06 and 0.14 with a 0.1 mean. *GW* estimates are calculated using the mean value of the spatially averaged specific yield while the range of possible values was used to estimate the uncertainty associated with these estimates.

2.4. Soil Moisture

[14] Where a sufficient monitoring network exists, a spatially averaged time series of soil moisture can be obtained from field observations [*Swenson et al.*, 2006]. In regions of sparse field data, an alternative is to use results from global model simulations [*Rodell et al.*, 2007; *Strassberg et al.*, 2007]. In the Murray Darling Basin, we derived monthly soil moisture storage values for the basin (*SM*) from January 2000 to January 2008 from the NOAH land surface model [*Ek et al.*, 2003], with the NOAH simulations being driven (parameterization and forcing) by the Global Land Data Assimilation System [*Rodell et al.*, 2004b]. The NOAH model simulates surface energy and water fluxes/budgets (including soil moisture) in response to near-surface atmospheric forcing and depending on surface conditions (e.g., vegetation state, soil texture and slope) [*Ek et al.*, 2003]. The NOAH model outputs of soil moisture estimates have a 1° spatial resolution and, using four soil layers, are representative of the top 2 m of the soil.

2.5. Drought Severity

[15] The severity of a hydrological drought can be determined by characterizing the deficit volume of water [*Tallaksen and Van Lanen*, 2004]. For a basin, the total water deficit $D(t)$ at time t below a drought threshold (φ_0) observed at time t_0 is

$$D(t) = [\varphi_0 - TWS(t)]I = [TWS(t_0) - TWS(t)]I \qquad (6)$$

where $D(t)$ is expressed in volume (km^3) or mm of equivalent water height, and I is an indicator function as follows:

$$I = \begin{cases} 1 & \text{if } TWS(t) \leq TWS(t_0) \\ 0 & \text{if } TWS(t) > TWS(t_0) \end{cases}$$

3. Results

[16] The analysis of the *SW* data shows that surface water resources, which provide most of the irrigation and domestic water supply for the ~2 million inhabitants of the region [*Kirby et al.*, 2006], are substantially affected by the multiyear drought. Figure 2 shows that, in as little as 2.4 years, *SW* in the basin declined by 83% from 19 km^3 in November 2000 (~76% of total infrastructure capacity) to 3.2 km^3 in April 2003 (~13% of total infrastructure capacity). From August 2002 onward the total surface water storage (*SW*) remains below 11.4 km^3 (~45% of total infrastructure capacity). Severe water restrictions prevented the complete drying up of the surface water resources which reached a record low since comprehensive monitoring began (circa 1980) of 1.8 km^3 in April 2007 (~7% of total infrastructure capacity). A multiannual drying trend in the GRACE *TWS* data (Figure 3a) reveals a more substantial loss of water in the basin than that observed in the *SW* data alone (Figure 3d). GRACE *TWS* shows an accumulated reduction of ~130 mm equivalent water depth between August 2002 and December 2006, an estimated total water loss across the basin of ~140 km^3 (Figure 3a).

[17] It is interesting to place the water deficits in a global climate change context. The 522 km^3 total rainfall deficit between 2001 and 2006 equates to a 1.5 mm increase in equivalent global sea level, or ~0.25 mm a^{-1}. This is comparable to half the contribution rate from the melting of the Greenland ice sheet [*Meier et al.*, 2007]. The ~120 km^3 *TWS* increase from the peak of the drought in March 2007 to January 2008 in the Murray-Darling Basin equates to a global sea level change of −0.34 mm. This balances the annual contribution of the melting of West Antarctica to the global sea level rise during 2006 [*Meier et al.*, 2007].

[18] GLDAS-NOAH simulations of soil moisture (*SM*) range from 5 to 29% (in volumetric water content) across the basin for the study period, and are within typical values for monthly means at 1° resolution [*Lawrence and Hornberger*, 2007]. A rapid decline in soil moisture is noticed at the start of the drought. *SM* values indicate a 120 km^3 loss in soil moisture storage in the 2 years after November 2000, with small subsequent fluctuations. For the 2002–2007 period, spatially averaged monthly soil moisture anomalies across the basin vary from a minimum of −44 km^3 in December 2006 to a maximum of 22 km^3 in July 2005 (relative to the 2000–2007 mean; Figure 3c), with a linear trend of −2 km^3 a^{-1}. Soil moisture declines rapidly during a drought and tends to stabilize once the capacity of the soil to dry out has been reached [*Yeh and Famiglietti*, 2008].

[19] A loss of groundwater occurred every year between 2001 and 2007 (Figure 3b) and the total groundwater loss between 2001 and 2007 is estimated at ~104 ± 40 km^3. Since the onset of the drought in 2001, the greatest decline in *GW* storage occurred between 2001 and 2003 (mean $GW_{2003} - GW_{2001} = -45$ km^3; Figure 3b). However, groundwater storage continues to decline thereafter, with a substantial loss of 59 km^3 observed between 2003 and 2007 (mean $GW_{2007} - GW_{2003} = -59$ km^3; Figure 3b). An increase in the delay and persistence of droughts as they propagate through the hydrological cycle to groundwater levels has been reported in temperate regions [*Eltahir and Yeh*, 1999; *Peters et al.*, 2003; *Peters et al.*, 2006] but has not previously been observed on such a long time scale. While the surface water and soil moisture droughts stabilized at low levels about 2 years after the onset of the drought, groundwater levels continued to decline significantly for the next 5 years.

[20] Between 2003 and 2007 the estimated linear rate of water loss for the GRACE *TWS* time series is comparable to that observed for the annual total water storage from in situ observations and modeling (*GW* + *SW* + *SM*) data. The combined annual anomalies of surface water, groundwater and soil moisture are highly correlated with the annual GRACE *TWS* (R = 0.94 and mean absolute difference = 13 km^3 for the 2003–2007 period; Figure 4a). Correlation between mean annual GRACE *TWS* − (*SW* + *SM*) and annual *GW* is also high ($R = 0.92$ for the 2003–2007 period; Figure 4b). Significant correlation has been found previously for seasonal fluctuations between GRACE data and groundwater from the saturated (*GW*) and unsaturated zone (*SM*) in the semiarid High Plains aquifer, USA [*Strassberg et al.*, 2007]. Our results confirm that GRACE can also detect long-term groundwater trends.

[21] A significant decline in the *TWS* is observed between 2002 and 2006. For this period the total water losses in annual soil moisture and groundwater are 9 km^3 and 56 km^3, respectively. Changes in total surface water storage are relatively small compared to the other major water stores in the basin (*GW* and *SM*). A minor gain of ~3 km^3 in the mean annual volume of surface water is observed between 2003 and 2005, while the total surface water loss between 2002 and 2006 is 2 km^3 (Figure 3d). Of the total water lost between 2002 and 2006, ~83% is groundwater, 14% is soil moisture and only 3% is surface water indicating that groundwater loss accounts for most of the GRACE-observed *TWS* loss. Aquifers often represent the largest water store in semiarid regions [*Simmers*, 2003], and our results highlight the importance of accounting for groundwater when assessing droughts and the long-term impact of environmental change on water resources.

[22] An increase in the GRACE *TWS* is noticed during the 2007–2008 period. We attribute this *TWS* increase to a return

Figure 3. Change of water storage in the main water stores of the Murray-Darling Basin during the multiyear drought. (a) Total water storage (*TWS*) anomalies relative to the mean from August 2002 to April 2008 estimated from GRACE solutions at 10-day epochs. The 1 sigma uncertainties are shown. (b) Groundwater storage (*GW*) anomalies relative to the mean for the 2000–2007 period. Vertical bars represent the range of *GW* estimates based on the spread of possible specific yield values. (c) Soil moisture storage (*SM*) anomalies relative to the mean for the 2000–2008 period and modeled from GLDAS-NOAH. (d) Surface water storage (*SW*) anomalies relative to the mean for the 2000–2008 period and estimated from the major lakes, reservoirs, weirs, and in-channel storages across the basin.

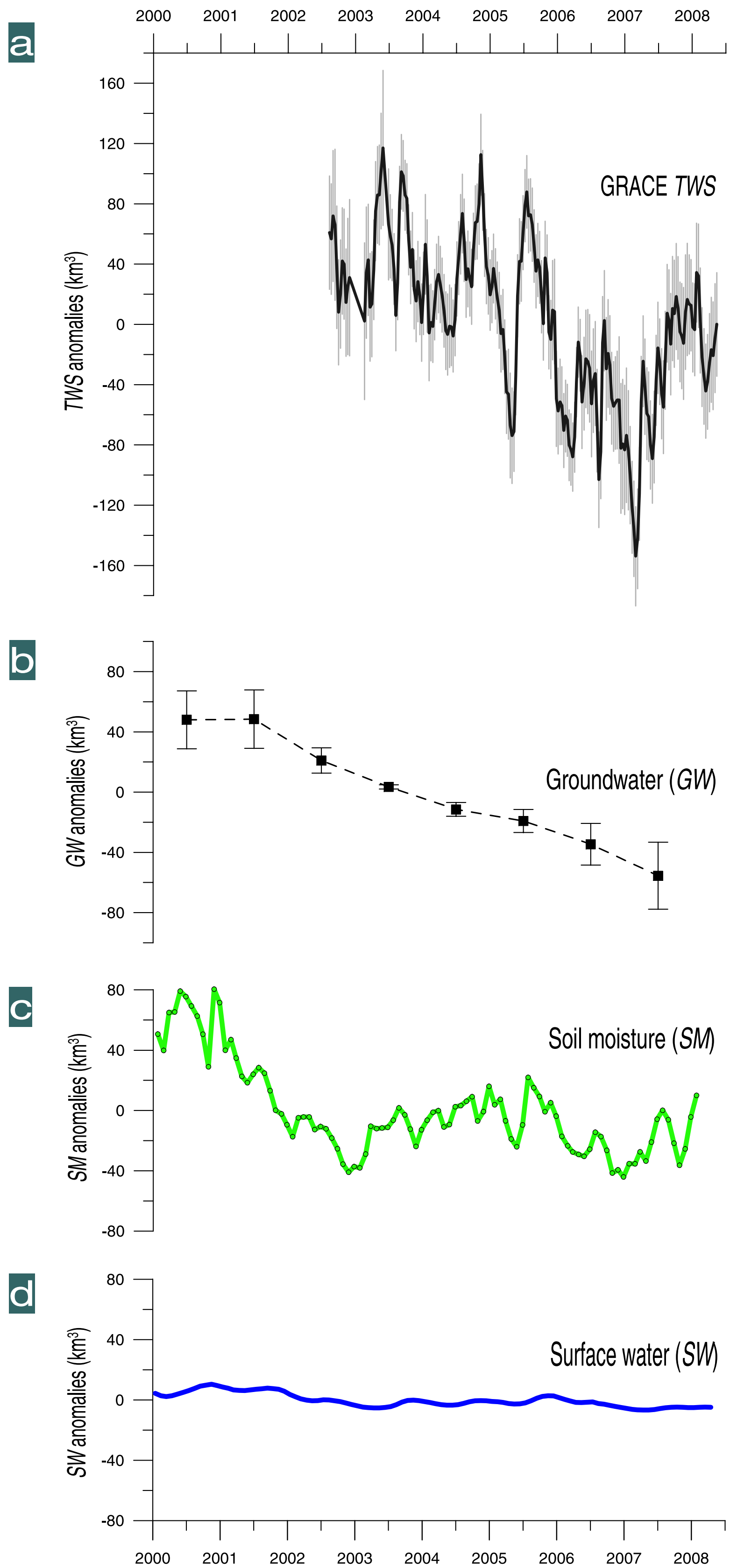
a
2000
2001
2002
2003
2004
2005
2006
2007
2008
GRACE TWS
TWS anomalies (km³)
b
Groundwater (GW)
GW anomalies (km³)
c
Soil moisture (SM)
SM anomalies (km³)
d
Surface water (SW)
SW anomalies (km³)

Figure 3

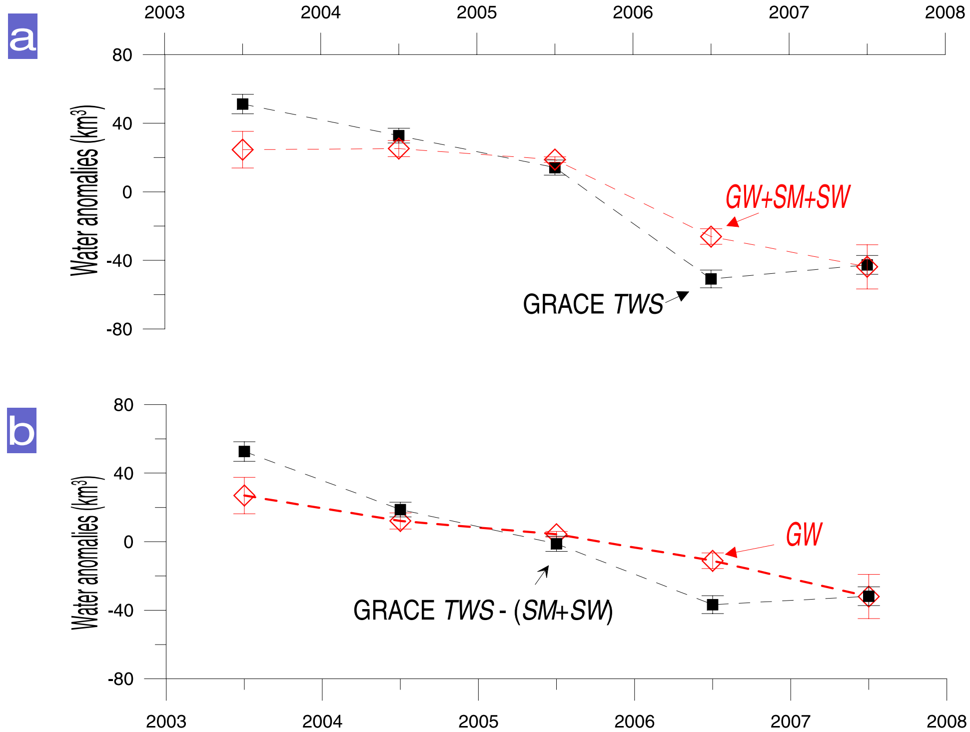

Figure 4. Comparison of GRACE *TWS* anomalies with hydrological estimates from in situ measurements (*SW* and *GW*) and modeling (*SM*). Vertical error bars indicate the uncertainties in *TWS* (black) and *GW* (red) estimates. (a) Comparison of annual GRACE *TWS* anomalies with annual combined anomalies from *GW*, *SM*, and *SW* for the 2003–2007 period. (b) Comparison of annual *GW* anomalies with annual combined anomalies from GRACE *TWS* − (*SM* + *SW*).

to average rainfall conditions from January to June 2007 and above average rainfall conditions between November 2007 and February 2008 when a large flooding event occurred in the northern part of the basin. Elsewhere, there has been limited runoff associated with these rainfall events ($\sim$2 km^3 *SW* increase from the lowest storage levels in April 2007 to March 2008). It is commonly observed that during a drought most of the rainwater is first used to replenish storage in the dry soil [e.g., *Eltahir and Yeh*, 1999]. Simulated *SM* values indicate that between January 2007 and January 2008, ΔSM_{2007} (45 km^3) accounts for $\sim$47% of the ΔTWS_{2007} ($\sim$95 km^3) increase. $\Delta TWS_{2007} - \Delta(SM + SW)_{2007}$ indicates $\sim$50 km^3 of water was thus potentially available for recharging of groundwater. However, because of the time lag for groundwater recharge to occur [*Scanlon et al.*, 2006], it is too early to confirm if these rainfall events have had an impact on groundwater levels.

4. Drought Indicators

[23] A hydrological drought is commonly defined as a deficit in surface water and groundwater [*Wilhite*, 2000]. Quantitative analysis of hydrological droughts is possible using temporal and volume characteristics of the water deficit and the threshold level method is the most frequently applied [*Tallaksen and Van Lanen*, 2004]. The duration of a hydrological drought can be defined as the time during which a hydrological variable (or a combination of variables) is consistently below a threshold level, while the severity is the volume of the deficit below this threshold. GRACE *TWS* anomalies include soil moisture (generally excluded from the definition of a hydrologic drought) and therefore may not be strictly used for the quantification and monitoring of hydrological droughts according to the definition above; however, GRACE *TWS* data allow the direct monitoring of the total water deficit across a large basin at 10-day time intervals and thus provides a valuable tool for studying hydrological droughts. Soil moisture is a major water store in most basins, which influences surface water storage via runoff and contributes to groundwater storage via deep infiltration. It is therefore important to include it for comprehensive water accounting.

[24] Once a drought threshold is determined, the total water storage deficit below this threshold can be used as a measure of the drought severity. The clear detection of the drying trend in the GRACE *TWS* data shows that GRACE data may be used to define the onset and end of a drought. However, in the case of the Murray-Darling Basin, the multiyear drought started before and continues beyond the GRACE time series currently available. 2001 corresponds to the start of the rainfall deficit at basin scale and to the onset of the decline in surface water and soil moisture storages (Figures 3c and 3d). Therefore, we take the mean ($SW + GW + SM$) in 2001 as the drought threshold level in order to calculate subsequent drought characteristics with $(SW + GW + SM)_{2001} - TWS_{2003} \sim (SW + GW + SM)_{2001} - (SW + GW + SM)_{2003} = 90$ km^3. Thus, across the Murray-Darling Basin the average drought severity (average total deficit volume) is calculated as $\sim$140 km^3 for the GRACE study period, with a maximum severity for 3 consecutive months of $\sim$240 km^3 observed between January and March 2007 (Figure 5).

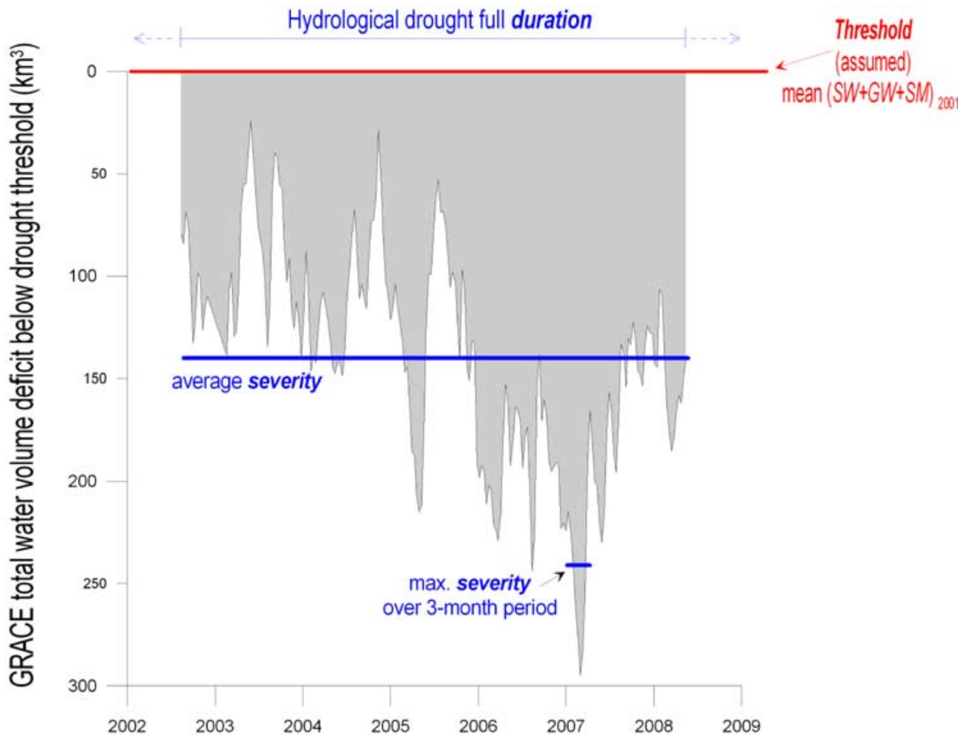

Figure 5. Severity of the multiyear drought derived from GRACE total water deficit across the Murray-Darling Basin.

[25] Drought severity can also be considered from the perspective of human demand for water supply [*Wilhite*, 2000; *Tallaksen and Van Lanen*, 2004]. Under "normal" conditions in the Murray-Darling Basin the annual average surface water use across the basin is ~11 km^3 (maximum storage capacity of ~25 km^3), while Australian state water agencies estimated the sustainable groundwater yield in 2002–2003 as ~2.4 km^3 a^{-1} [*Kirby et al.*, 2006]. GRACE observations indicate a total water deficit of ~140 km^3 between August 2002 and December 2006, that is, ~-32 km^3 a^{-1} or about 2.4 times the annual rate of average water consumption across the basin under "normal" conditions.

5. Discussion

[26] Drought is a recurring issue in many parts of Australia including the Murray-Darling Basin and it can be argued that many natural ecosystems are adapted to cope with this variability [*McMahon and Finlayson*, 2003]. However, government agencies are reporting unprecedented socioeconomical and environmental impacts: significant loss of aquatic and riparian flora and fauna, lowest water supply storage on record implying severe water restrictions for urban centers and irrigation, significant decrease in agricultural production and increase in fire risk both in terms of intensity and season length [*Murphy and Timbal*, 2008]. Concern over anthropogenically driven climate change has placed the need for improvement in our understanding of the impact of climate change on global water resources at the center of the international research agenda [*Intergovernmental Panel on Climate Change*, 2007; *Milly et al.*, 2008]. It is important to establish whether the severe, ongoing drought in the Murray-Darling Basin is only part of the natural variability or is linked to anthropogenic climate change. Enhanced greenhouse gas concentrations are likely to be an influence on increasingly arid conditions in the Murray-Darling Basin, at least on rising temperature which exacerbates the dry conditions during the ongoing drought, a phenomenon not observed during previous prolonged droughts [*Murphy and Timbal*, 2008]. Climates models also indicate that average rainfall (especially the first half of winter) in the southern part of the Murray-Darling Basin is likely to decline in the future as greenhouse gas concentrations increase [*Timbal and Jones*, 2008]. Our observations of severe water shortages therefore give an indication of potential future water stresses to expect according to climate model scenarios for the southern half of the Murray-Darling Basin [*Murphy and Timbal*, 2008; *Timbal and Jones*, 2008] and confirm the urgent need to reconsider the planning of water resources management and infrastructure under a changing climate [*Milly et al.*, 2008].

[27] Global environmental changes driven by human activities also include land use change. The respective impact of land use change and multiyear drought on water resources can be complex and converse. In a large region of West Africa in the Sahel it was found that, despite the long-lasting droughts of the 1970s and 1980s, surface and groundwater resources increased following intensive land clearing [*Leduc et al.*, 2001; *Leblanc et al.*, 2008]. Land clearing in the Murray-Darling Basin which started with European settlement in early 1800s caused a general rise of the water table still observed during the second half of the 20th century, and subsequently led to the appearance of dryland salinity [*Allison et al.*, 1990]. The observed decline of shallow groundwater levels and GRACE TWS show that the ongoing multiyear drought in the Murray-Darling Basin has, at least temporarily, reversed the long-term groundwater trend inherited from land clear-

ance and may induce a temporary halt of secondary salinity processes.

[28] The propagation of a drought into the various reservoirs of the hydrological cycle (soil moisture, surface water and groundwater) can be asynchronous and of various magnitude, with groundwater generally having the longest time lag [*Eltahir and Yeh*, 1999; *Wilhite*, 2000; *Tallaksen and Van Lanen*, 2004]. This may be further exacerbated in arid and semiarid areas where groundwater recharge occurs at time scales varying from weeks to decades, depending partly on the recharge processes and rates and the thickness of the unsaturated zone [*Scanlon et al.*, 2006]. Extreme dry events (droughts) have a more persistent effect on groundwater levels than extreme wet events (floods) because of the nonlinear dependence of the groundwater discharge on groundwater levels [*Eltahir and Yeh*, 1999; *Peters et al.*, 2006]. The effects of a drought on groundwater resources include the decline of groundwater levels and the decrease of groundwater discharge to springs, surface water bodies and riparian zones [*Peters et al.*, 2003]. Across the basin, shallow aquifers potentially perform important ecological services to groundwater-dependent ecosystems [*Murray et al.*, 2003; *Eamus et al.*, 2006] on which the drought is impacting. Where groundwater salinity is low, shallow aquifers are also used for water consumption [*Ife and Skelt*, 2004]. Changes in groundwater storage may also have indirect feedbacks on the climate. Shallow water tables can influence soil moisture and evapotranspiration [*Chen and Hu*, 2004; *Miguez-Macho et al.*, 2007] while in areas populated by deep-rooted trees deeper water tables can influence transpiration. These phenomena explain why groundwater influences the energy budget and water balances at the land surface and why groundwater droughts can have feedback mechanisms on the climate. Our observations in the Murray-Darling Basin therefore confirm that it is important to account for groundwater in climate models [e.g., *Niu et al.*, 2007; *Miguez-Macho et al.*, 2007] and that, in regions with sparse groundwater monitoring, GRACE *TWS* estimates provide useful basin-scale estimates [*Niu et al.*, 2007].

6. Conclusions

[29] GRACE provides integrated observations for quantifying the severity and characterizing drought at the basin scale. The long GRACE time series now available allows the tracking of total water deficit over a long period and the monitoring of multiyear droughts.

[30] The combination of terrestrial hydrologic and space gravity observations shows that the drought conditions in the Murray-Darling Basin in southeast Australia began around 2001, with over 50% of water loss occurring in groundwater resources. Since 2001, the average annual groundwater loss amounts to $\sim$17 km^3 or about 7 times the annual sustainable yield adopted for groundwater usage. The average annual loss of surface water and groundwater amounts to 20 km^3, or nearly 150% of the total water usage under "normal" conditions. This shows the significant vulnerability of agriculture in the basin.

[31] In 2007 and early 2008, the meteorological drought in the Murray-Darling Basin abated with a return to average or above average monthly rainfall; however, the GRACE *TWS* data show that a substantial accumulated bulk water deficit remains in the basin. Since March 2008, rainfall has returned to deficit conditions (last rainfall data available June 2008).

[32] **Acknowledgments.** This work was partly supported by the Australian Research Council grant DP0666300. Access to in situ surface water and groundwater data was facilitated by the Murray-Darling Basin Commission, the Queensland Department of Natural Resources and Water, the NSW Department of Water and Energy, and the Victorian Department of Sustainability and Environment. GLDAS data are from the NASA Goddard Space Flight Center. We thank the associate editor and three anonymous reviewers for their constructive comments.

References

Allison, G. B., P. G. Cook, S. R. Barnett, G. R. Walker, I. D. Jolly, and M. W. Hughes (1990), Land clearance and river salinisation in the western Murray Basin, Australia, *J. Hydrol.*, *119*, 1–20, doi:10.1016/0022-1694(90)90030-2.

Andersen, O. B., S. I. Seneviratne, J. Hinderer, and P. Viterbo (2005), GRACE-derived terrestrial water storage depletion associated with the 2003 European heat wave, *Geophys. Res. Lett.*, *32*, L18405, doi:10.1029/2005GL023574.

Bryant, E. (2005), *Natural Hazards*, Cambridge Univ. Press, Cambridge, U. K.

Carrère, L., and F. Lyard (2003), Modeling of the barotropic response of the global ocean to atmospheric wind and pressure forcing—Comparison with observations, *Geophys. Res. Lett.*, *30*(6), 1275, doi:10.1029/2002GL016473.

Chen, X., and Q. Hu (2004), Groundwater influences on soil moisture and surface evaporation, *J. Hydrol.*, *297*, 285–300, doi:10.1016/j.jhydrol.2004.04.019.

Crowley, J. W., J. X. Mitrovica, R. C. Bailey, M. E. Tamisiea, and J. L. Davis (2006), Land water storage within the Congo Basin inferred from GRACE satellite gravity data, *Geophys. Res. Lett.*, *33*, L19402, doi:10.1029/2006GL027070.

Eamus, D., R. Froend, R. Loomes, G. Hose, and B. Murray (2006), A functional methodology for determining the groundwater regime needed to maintain the health of groundwater-dependent vegetation, *Aust. J. Bot.*, *54*(2), 97–114, doi:10.1071/BT05031.

Ek, M. B., K. E. Mitchell, Y. Lin, E. Rogers, P. Grunmann, V. Koren, G. Gayno, and J. D. Tarpley (2003), Implementation of Noah land surface model advances in the National Centers for Environmental Prediction operational mesoscale Eta model, *J. Geophys. Res.*, *108*(D22), 8851, doi:10.1029/2002JD003296.

Ellett, K. M., J. P. Walker, A. W. Western, and M. Rodell (2006), A framework for assessing the potential of remote-sensed gravity to provide new insight on the hydrology of the Murray-Darling Basin, *Aust. J. Water Res.*, *10*, 125–138.

Eltahir, E. A. B., and P. J.-F. Yeh (1999), On the asymmetric response of aquifer water level to floods and droughts in Illinois, *Water Resour. Res.*, *35*(4), 1199–1217, doi:10.1029/1998WR900071.

Heim, R. R. (2000), Drought indices: A review, in *Drought: A Global Assessment, Hazards Disasters Ser.*, vol. 1, edited by D. A. Wilhite, pp. 159–167, Routledge, London.

Ife, D., and K. Skelt (2004), Murray Darling Basin groundwater status 1990–2000, *Publ. 32/04*, Murray Darling Basin Comm., Canberra. (Available at http://www.mdbc.gov.au)

Intergovernmental Panel on Climate Change (2007), *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, edited by S. Solomon et al., Cambridge Univ. Press, Cambridge, U. K.

Fetter, C. W. (2001), *Applied Hydrogeology*, 4th ed., 598 pp., Prentice-Hall, Upper Saddle River, N. J.

Johnson, A. I. (1967), Specific yield-Compilation of specific yields for various materials, 74 pp., *U.S. Geol. Surv. Water Supply Pap., 1662-D*.

Kirby, M., R. Evans, G. Walker, R. Cresswell, J. Coram, S. Khan, Z. Paydar, M. Mainuddin, N. McKenzie, and S. Ryan (2006), The shared water resources of the Murray-Darling Basin, *Publ. 21/06*, Murray-Darling Basin Comm., Canberra. (Available at http://www.mdbc.gov.au)

Keyantash, J., and J. A. Dracup (2002), The quantification of drought: An evaluation of drought indices, *Bull. Am. Meteorol. Soc.*, *83*, 1167–1180.

Keyantash, J. A., and J. A. Dracup (2004), An aggregate drought index: Assessing drought severity based on fluctuations in the hydrologic cycle and surface water storage, *Water Resour. Res.*, *40*, W09304, doi:10.1029/2003WR002610.

Lawrence, J. E., and G. M. Hornberger (2007), Soil moisture variability across climate zones, *Geophys. Res. Lett.*, *34*, L20402, doi:10.1029/2007GL031382.

Leblanc, M., G. Favreau, S. Massuel, S. Tweed, M. Loireau, and B. Cappelaere (2008), Land clearance and hydrological change in the Sahel: SW Niger, *Global Planet. Change*, *61*, 135–150, doi:10.1016/j.gloplacha.2007.08.011.

Leduc, C., G. Favreau, and P. Schroeter (2001), Long-term rise in a Sahelian water table: The continental terminal in south-west Niger, *J. Hydrol.*, *243*, 43–54, doi:10.1016/S0022-1694(00)00403-0.

Lemoine, J.-M., S. Bruinsma, S. Loyer, R. Biancale, J.-C. Marty, F. Perosanz, and G. Balmino (2007), Temporal gravity field models inferred from GRACE data, *Adv. Space Res.*, *39*, 1620–1629, doi:10.1016/j.asr.2007.03.062.

Le Provost, C., F. Lyard, M. L. Genco, and F. Rabilloud (1998), A hydrodynamic ocean tide model improved by assimilation of satellite altimeter-derived data set, *J. Geophys. Res.*, *103*(C3), 5513–5529, doi:10.1029/97JC01733.

McMahon, T. A., and B. L. Finlayson (2003), Droughts and anti-droughts: The low flow hydrology of Australian rivers, *Freshwater Biol.*, *48*, 1147–1160, doi:10.1046/j.1365-2427.2003.01098.x.

Meier, M. F., M. B. Dyurgerov, U. K. Rick, S. O'Neel, W. T. Pfeffer, R. S. Anderson, S. P. Anderson, and A. F. Glazovsky (2007), Glaciers dominate eustatic sea-level rise in the 21st century, *Science*, *317*, 1064–1067, doi:10.1126/science.1143906.

Miguez-Macho, G., Y. Fan, C. P. Weaver, R. Walko, and A. Robock (2007), Incorporating water table dynamics in climate modeling: 2. Formulation, validation, and soil moisture simulation, *J. Geophys. Res.*, *112*, D13108, doi:10.1029/2006JD008112.

Milly, P. C. D., J. Betancourt, M. Falkenmark, R. M. Hirsch, Z. W. Kundzewicz, D. P. Lettenmaier, and R. J. Stouffer (2008), Stationarity is dead: Whither water management?, *Science*, *319*, 573–574, doi:10.1126/science.1151915.

Murphy, B., and B. Timbal (2008), A review of recent climate variability and climate change in south eastern Australia, *Int. J. Climatol.*, *28*, 859–879, doi:10.1002/joc.1627.

Murray, B. R., M. J. B. Zeppel, G. C. Hose, and D. Eamus (2003), Groundwater-dependent ecosystems in Australia: It's more than just water for rivers, *Ecol. Manage. Restor.*, *4*(2), 110–113, doi:10.1046/j.1442-8903.2003.00144.x.

Niu, G.-Y., Z.-L. Yang, R. E. Dickinson, L. E. Gulden, and H. Su (2007), Development of a simple groundwater model for use in climate models and evaluation with Gravity Recovery and Climate Experiment data, *J. Geophys. Res.*, *112*, D07103, doi:10.1029/2006JD007522.

Peters, E., P. J. Torfs, H. A. Van Lanen, and G. Bier (2003), Propagation of drought through groundwater—A new approach using linear reservoir theory, *Hydrol. Processes*, *17*, 3023–3040, doi:10.1002/hyp.1274.

Peters, E., G. Bier, H. A. Van Lanen, and P. J. Torfs (2006), Propagation and spatial distribution of drought in a groundwater catchment, *J. Hydrol.*, *321*, 257–275, doi:10.1016/j.jhydrol.2005.08.004.

Petheram, C., W. Dawes, G. Walker, and R. B. Grayson (2003), Testing in-class variability of groundwater systems: Local upland systems, *Hydrol. Processes*, *17*, 2297–2313, doi:10.1002/hyp.1333.

Ramillien, G., A. Cazenave, and O. Brunau (2004), Global time variations of hydrological signals from GRACE satellite gravimetry, *Geophys. J. Int.*, *158*(3), 813–826, doi:10.1111/j.1365-246X.2004.02328.x.

Ramillien, G., F. Frappart, A. Cazenave, and A. Guntner (2005), Time variations of land water storage from an inversion of 2 years of GRACE geoids, *Earth Planet. Sci. Lett.*, *235*, 283–301, doi:10.1016/j.epsl.2005.04.005.

Ramillien, G., A. Lombard, A. Cazenave, E. R. Ivins, M. Llubes, F. Remy, and R. Biancale (2006), Interannual variations of the mass balance of the Antarctica and Greenland ice sheets from GRACE, *Global Planet. Change*, *53*, 198–208, doi:10.1016/j.gloplacha.2006.06.003.

Ritchie, J. W., G. Y. Abawi, S. C. Dutta, T. R. Harris, and M. Bange (2004), Risk management strategies using seasonal climate forecasting in irrigated cotton production: A tale of stochastic dominance, *Aust. J. Agric. Resour. Econ.*, *48*(1), 65–93, doi:10.1111/j.1467-8489.2004.t01-1-00230.x.

Rodell, M., and J. S. Famiglietti (1999), Detectability of variations in continental water storage from satellite observations of the time dependent gravity field, *Water Resour. Res.*, *35*(9), 2705–2723, doi:10.1029/1999WR900141.

Rodell, M., J. S. Famiglietti, J. Chen, S. I. Seneviratne, P. Viterbo, S. Holl, and C. R. Wilson (2004a), Basin scale estimates of evapotranspiration using GRACE and other observations, *Geophys. Res. Lett.*, *31*, L20504, doi:10.1029/2004GL020873.

Rodell, M., P. R. Houser, U. Jambor, and J. Gottschalck (2004b), The Global Land Data Assimilation System, *Bull. Am. Meteorol. Soc.*, *85*(3), 381–394, doi:10.1175/BAMS-85-3-381.

Rodell, M., J. Chen, H. Kato, J. Famiglietti, J. Nigro, and C. Wilson (2007), Estimating groundwater storage changes in the Mississippi River basin (USA) using GRACE, *Hydrogeol. J.*, *15*(1), 159–166, doi:10.1007/s10040-006-0103-7.

Scanlon, B. R., K. E. Keese, A. L. Flint, L. E. Flint, C. B. Gaye, W. M. Edmunds, and I. Simmers (2006), Global synthesis of groundwater recharge in semiarid and arid regions, *Hydrol. Processes*, *20*, 3335–3370, doi:10.1002/hyp.6335.

Schmidt, R., et al. (2006), GRACE observations of changes in continental water storage, *Global Planet. Change*, *50*, 112–126, doi:10.1016/j.gloplacha.2004.11.018.

Simmers, I. (2003), *Understanding Water in a Dry Environment: Hydrological Processes in Arid and Semi-arid Zones*, *Int. Contrib. Hydrogeol.*, vol. 23, Balkema, Lisse, Netherlands.

Strassberg, G., B. R. Scanlon, and M. Rodell (2007), Comparison of seasonal terrestrial water storage variations from GRACE with groundwater-level measurements from the High Plains Aquifer (USA), *Geophys. Res. Lett.*, *34*, L14402, doi:10.1029/2007GL030139.

Swenson, S. C., P. J.-F. Yeh, J. Wahr, and J. S. Famiglietti (2006), A comparison of terrestrial water storage variations from GRACE with in situ measurements from Illinois, *Geophys. Res. Lett.*, *33*, L16401, doi:10.1029/2006GL026962.

Syed, T. H., J. S. Famiglietti, M. Rodell, J. Chen, and C. R. Wilson (2008), Analysis of terrestrial water storage changes from GRACE and GLDAS, *Water Resour. Res.*, *44*, W02433, doi:10.1029/2006WR005779.

Tallaksen, L. M., and H. A. Van Lanen (Eds.) (2004), *Hydrological Drought: Processes and Estimation Methods for Streamflow and Groundwater*, *Dev. Water Sci.*, vol. 48, 579 pp., Elsevier, Amsterdam.

Timbal, B., and D. Jones (2008), Future projections of winter rainfall in southeast Australia using a statistical downscaling technique, *Clim. Change*, *86*, 165–187, doi:10.1007/s10584-007-9279-7.

Van Dijk, A., G. Podger, and M. Kirby (2007), Integrated water resources management in the Murray-Darling Basin: Increasing demands on decreasing supplies, in *Reducing the Vulnerability of Societies to Water Related Risks at the Basin Scale*, edited by A. Schumann and M. Pahlow, IAHS Publ., *317*, pp. 24–30.

Wahr, J., S. Swenson, V. Zlotnicki, and I. Velicogna (2004), Time-variable gravity from GRACE: First results, *Geophys. Res. Lett.*, *31*, L11501, doi:10.1029/2004GL019779.

Wilhite, D. A. (Ed.) (2000), *Drought: A Global Assessment*, *Hazards Disasters Ser.*, Routledge, London.

Wilhite, D. A., and M. H. Glantz (1985), Understanding the drought phenomenon: the role of definitions, *Water Int.*, *10*(3), 111–120.

Yeh, P. J.-F., and J. S. Famiglietti (2008), Regional terrestrial water storage change and evapotranspiration from terrestrial and atmospheric water balance computations, *J. Geophys. Res.*, *113*, D09108, doi:10.1029/2007JD009045.

A. Fakes, CSIRO Sustainable Ecosystems, James Cook University, Cairns, Qld 4870, Australia.

M. J. Leblanc and S. O. Tweed, Hydrological Sciences Research Unit, School of Earth and Environmental Sciences, James Cook University, PO Box 6811, Cairns, Qld 4870, Australia. (marc.leblanc@jcu.edu.au)

G. Ramillien, Dynamique Terrestre et Planétaire, Observatoire Midi-Pyrénées, UMR5562, CNRS, F-31400, Toulouse, France.

P. Tregoning, Research School of Earth Sciences, Australian National University, Canberra, ACT 0200, Australia.