Assessing a land surface model's improvements with GRACE estimates

Guo-Yue Niu[1] and Zong-Liang Yang[1]

Received 20 December 2005; revised 7 February 2006; accepted 1 March 2006; published 6 April 2006.

[1] The Gravity Recovery and Climate Experiment (GRACE) satellites have produced an unprecedented data set of terrestrial water storage (TWS) change in large-scale river basins. Recent research has found that monthly variations of soil moisture and snow water simulated by land surface models compared favorably with the GRACE-derived TWS change. Compared to the GRACE data, the standard version of the National Center for Atmospheric Research (NCAR) Community Land Model (CLM) produces a weaker TWS variability in tropical and mid-latitudes but a stronger TWS variation in high latitudes. However, a modified version of CLM that includes more realistic interception, runoff, and frozen soil processes improves the simulation of TWS change in global river basins of various scales. In addition, the modified CLM improves the modeling of evapotranspiration through the improvements in the modeling of TWS variation and runoff in the Amazon River basin. Along this line of research, this paper shows that such GRACE data can be used as a means of evaluating the hydrological schemes in a land surface model. **Citation:** Niu, G.-Y., and Z.-L. Yang (2006), Assessing a land surface model's improvements with GRACE estimates, *Geophys. Res. Lett.*, *33*, L07401, doi:10.1029/2005GL025555.

1. Introduction

[2] Terrestrial water storage (TWS) includes soil moisture, snow and ice, groundwater, lakes and rivers, and water contained in biomass. TWS change reflects the change in fresh water resources that sustains various life forms. TWS change is also an indicator of Earth's climate variability. On the other hand, changes in TWS can feed back to affect various aspects of Earth's hydrological cycle. Numerous studies have shown that soil moisture conditions influence surface energy and water balances, which in turn affect weather and climate over multiple scales [*Koster et al.*, 2004]. It is also extensively documented in the literature that snow cover has a significant impact on soil moisture and precipitation [*Bamzai and Shukla*, 1999]. Recent studies show groundwater can also have a large impact on surface energy and water balances in regions where the water table is shallow [e.g., *Gutowski et al.*, 2002].

[3] The Gravity Recovery And Climate Experiment (GRACE) satellites, launched March 17, 2002, are measuring Earth's gravity field with enough precision to infer TWS change (ΔS) over sufficiently large regions [*Wahr et al.*, 2004; *Tapley et al.*, 2004]. *Rodell et al.* [2004a] demonstrated that the GRACE-derived ΔS is useful to estimate basin-scale evapotranspiration (*ET*) when combined with observed precipitation and river discharge data. The GRACE-derived ΔS can be also applied to evaluate the simulations of snowmelt runoff and infiltration through frozen ground in cold regions [*Niu and Yang*, 2006].

[4] The GRACE-derived ΔS may facilitate improving land surface models for use in climate models. The modeled *ET*, which is a key variable to couple land surface processes with atmospheric processes, is difficult to evaluate at a regional scale because of a lack of observed *ET* at such a scale. The water balance for a region or a river basin is:

$$ET = P - R - \Delta S \tag{1}$$

where P is precipitation, R is total runoff, and ΔS is the TWS change. Driven by the observed precipitation (and other forcing data), a land surface model can produce *ET*, R, and ΔS. The modeled *ET* can be indirectly evaluated when R and ΔS are validated against river discharge and the GRACE-derived ΔS, respectively.

[5] Most land surface models are confined to a certain depth of soil (e.g., the total soil depth of the NCAR CLM is globally 3.43 m), thereby excluding the groundwater variations in regions where the water table is deep. In addition, the water storage in lakes and rivers is usually not explicitly represented in these models. Researchers [*Wahr et al.*, 2004; *Tapley et al.*, 2004; *Chen et al.*, 2005] usually evaluate the GRACE-derived ΔS against those modeled by such a land model (e.g., the GLDAS TWS [soil water and snow water storage] variations modeled by the Noah land model) in large-scale river basins for lack of ground-based observations at such a scale. It is still unknown how and to what extent a land surface model that excludes groundwater and water storage in lakes and rivers can affect the TWS seasonal variability. Moreover, how various representations of terrestrial hydrological processes affect the modeling of TWS variations with such a land model is unknown. We address the following questions by employing the National Center for Atmospheric Research (NCAR) Community Land Model (CLM): (1) How do different model designs affect the simulation of the TWS seasonal variability? and (2) Can GRACE-derived TWS variations help improve the modeling of *ET* over a region or a river basin?

2. Model

[6] The NCAR CLM has been developed to represent the terrestrial thermal and hydrological processes in a fully coupled climate system model. *Oleson et al.* [2004] provided a thorough description of the standard CLM. The terrestrial hydrological processes in the model include interception of precipitation by the vegetation canopy, snow accumulation and ablation, infiltration and percolation of soil water, freezing and thawing of soil water, surface runoff

[1]Department of Geological Sciences, John A. and Katherine G. Jackson School of Geosciences, University of Texas at Austin, Austin, Texas, USA.

0094-8276/06/2005GL025555$05.00

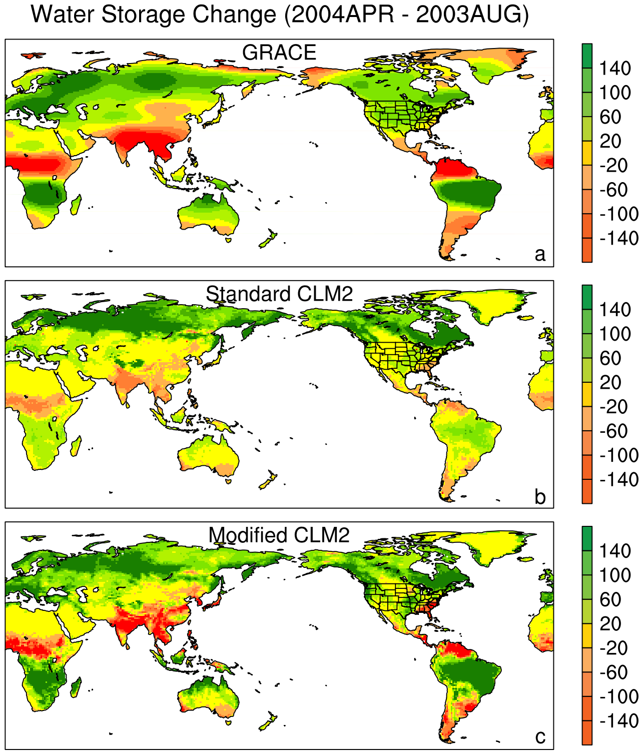

Figure 1. Changes in ΔS (mm) between April 2004 and August 2003 (April 2004 minus August 2003). (a) GRACE by *Seo and Wilson* [2005], (b) the standard CLM, and (c) the modified CLM.

and base flow, and *ET*. Although water storage in lakes and rivers is not explicitly represented in the model, a lake model is included in the NCAR CLM to compute its water storage variations through the variations of precipitation and evaporation. The model has obvious shortcomings in representing the interception of precipitation by the vegetation canopy, surface runoff, subsurface runoff and frozen soil. *Niu et al.* [2005] and *Niu and Yang* [2006] proposed modifications to these schemes. The resulting version is referred to as the "modified CLM" in this paper.

2.1. Interception

[7] The standard CLM produces excessive interception of precipitation by the canopy and, in turn, results in excessive interception loss. The modified CLM implements a subgrid precipitation scheme to reduce the fractional area of the vegetation canopy that can receive precipitation and thus reduces the interception loss by a factor of three in the Amazon River basin. This modification allows more water to reach the ground.

2.2. Saturated Hydraulic Conductivity

[8] Following TOPMODEL, the standard CLM assumes the saturated hydraulic conductivity, K_{sat}, exponentially decays with soil depth but excludes surface macropores. The modified CLM defines K_{sat} as a function of soil texture (as climate models usually do) and reduces the gravitational drainage from the bottom of the soil column. This modification mainly enhances the subsurface layers' K_{sat} and thus the percolation rate.

2.3. Surface Runoff

[9] The standard CLM overestimates surface runoff by redundantly parameterizing surface runoff as a sum of the TOPMODEL-based surface runoff in the fractional saturated area and the BATS-type surface runoff in the fractional unsaturated area. *Niu et al.* [2005] found that the cumulative distribution function (CDF) of topographic indexes in a GCM grid cell decreases exponentially with the increase of topographic index for topographic indexes equal to and larger than the mean value of the topographic index of the grid cell. Therefore, the modified CLM represents the fractional saturated area as an exponential function of the water table depth. Snowmelt and rainfall in the fractional saturated area is immediately converted to surface runoff. This modification greatly decreases surface runoff and increases infiltration at the soil surface.

2.4. Subsurface Runoff (Base Flow)

[10] The standard CLM has the potential to generate an extremely large volume of base flow mainly due to the extremely large base flow coefficient, 4×10^{-2} mm s^{-1}, a value that exceeds any likely precipitation rates. Following TOPMODEL, the modified CLM parameterizes base flow as an exponential function of the water table depth with a significantly reduced base flow coefficient, 1.0×10^{-4} mm s^{-1}.

2.5. Frozen Soil

[11] In the standard CLM, frozen soil is extremely impermeable because it neglects the supercooled soil water (the liquid water that coexists with ice over a wide range of temperatures below 0°C) and its soil hydraulic properties are parameterized as functions of the extremely low liquid water content. *Niu and Yang* [2006] parameterized the supercooled soil water by applying the freezing-point depression equation. Additionally, a fractional permeable area in a GCM grid cell was proposed to increase the permeability of the frozen ground at a GCM grid scale.

[12] Note that the above modifications were not intentionally proposed to fit the GRACE-derived TWS variations. Justifications of the modifications are described in detail by *Niu et al.* [2005] and *Niu and Yang* [2006].

3. Experiments and Results

[13] We conducted two experiments: one with the standard CLM and one with the modified CLM. In both runs, the decay factor $f = 2.0$ m^{-1} [*Niu and Yang*, 2006]. We used the Global Land Data Assimilation System (GLDAS) $1° \times 1°$ 3-hourly, near-surface meteorological data for the years 2002–2004 [*Rodell et al.*, 2004b] to drive the model. The vegetation and soil parameters at $1° \times 1°$ were interpolated from the high-resolution raw data of the standard CLM version 2.0. To reduce the uncertainties induced by the initial conditions, we first ran the model for three years from 2002–2004 and saved the model prognostic variables at the end of the run. We then used the saved prognostic variables as the initial conditions for another three-year run.

[14] The change in the GRACE-derived ΔS between August 2003 and April 2004 (Figure 1a) remarkably shows the characteristics of ΔS changes in different climate zones (i.e., the cold, monsoon, and tropical rainforest regions). In

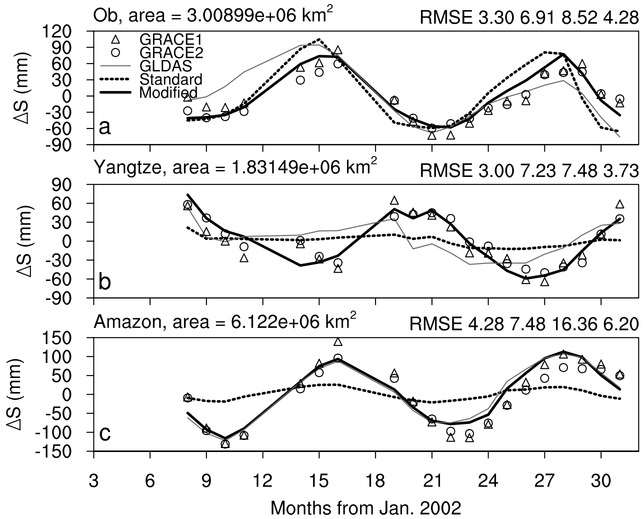

Figure 2. River-basin averaged water storage anomalies in (a) Ob, (b) Yangtze, and (c) Amazon river basins from GRACE (GRACE1 given by *Seo and Wilson* [2005], and GRACE2 given by *Chen et al.* [2005]), GLDAS, the standard CLM and the modified CLM. Also shown on the top-right of Figures 2a, 2b, and 2c are the root-mean-square-errors between each data set and GRACE1 in the order of GRACE2, GLDAS, Standard, and Modified.

cold regions, the change in ΔS is positive because of the snowpack that accumulated on the ground and the snowmelt water that infiltrated into the soil in April 2004. In monsoon regions (Africa, India, East Asia, and South America), the water storage shows negative changes, which is consistent with the drying after the monsoon-season precipitation in August 2003. In tropical rainforest regions (Amazon and Congo rivers), the change in ΔS is positive because August is a dry month. The standard CLM does a good job in cold regions except in Western Europe where the snowmelt water runs off immediately over a frozen ground. The standard CLM also produced smaller changes in monsoon and tropical rainforest regions (Figure 1b). However, the modified CLM produced water storage changes that compare more favorably with the GRACE-derived values in most of these regions (Figure 1c). Note that the model results should be smoothed as GRACE estimates to have a fair comparison. However, the large-scale water storage changes (e.g., in the Amazon basin) are so pronounced in both simulations that smoothing would not affect the above conclusions.

[15] We used two GRACE data sets that result from different filtering algorithms [*Seo and Wilson*, 2005; *Chen et al.*, 2005] to evaluate the modeled water storage variations. These two data sets contain 20 months starting from August 2002 to July 2004 with four months missing in between. We selected the same 20 months of the modeled data as those of GRACE to compute the seasonal variability of the water storage in Ob, Yangtze and Amazon River basins, which represent rivers in cold, monsoon, and tropical rainforest regions, respectively. The modified CLM better captures the seasonal variability of the water storage in all the three river basins than does the standard CLM (Figure 2). In the Ob River basin, the variability of water storage is affected by snow and frozen soil. The more permeable frozen ground in the modified CLM allows more snowmelt water to infiltrate into deeper soil and thus produces greater water storage in spring than does the standard CLM [*Niu and Yang*, 2006]. The modified CLM captures the variability of the GRACE-derived TWS change in most of the global 55 largest river basins (results not shown).

[16] The GRACE technique is more accurate for larger areas over longer times. So does the model because of the accuracy of the forcing data. Recent studies [*Wahr et al.*, 2004; *Tapley et al.*, 2004; *Chen et al.*, 2005] validated the GRACE estimates with the GLDAS water storage variations in the largest and "secondary" river basins. As shown in Figure 3, the modified CLM agrees fairly well with GRACE in some small basins at a scale around 300,000 km^2. This agreement not only indicates the modified CLM is superior to the standard CLM but also indicates that both GLDAS forcing data and GRACE data are sufficiently accurate for basins at such a scale.

[17] The GLDAS water storage variations are also plotted in Figures 2 and 3 as a reference; most researchers used this data set to evaluate the GRACE estimates. The root-mean-square-error (RMSE) between GLDAS and GRACE1 is 6.91 mm in Ob basin, while the RMSE between the modified CLM and GRACE1 is 4.28 mm, indicating that different model designs affect modeling TWS variations (most obviously in Ob, Yangtze, and Don). The RMSE between the standard CLM and the modified CLM in Ob is 5.21 mm, while the RMSE between the two GRACE estimates is 3.30 mm. The greater difference among model estimates than that between the two GRACE estimates implies that GRACE can be used to validate model's hydrological schemes.

[18] The modified CLM produces runoff in closer agreement with the GRDC data than does the standard CLM in the Amazon River basin (Figure 4a). Because of the more accurate R and ΔS (Figure 2c), the modified CLM is expected to produce more realistic ET constrained by the water balance equation (Equation (1); P is observed). ET from the modified CLM shows a weaker seasonality than

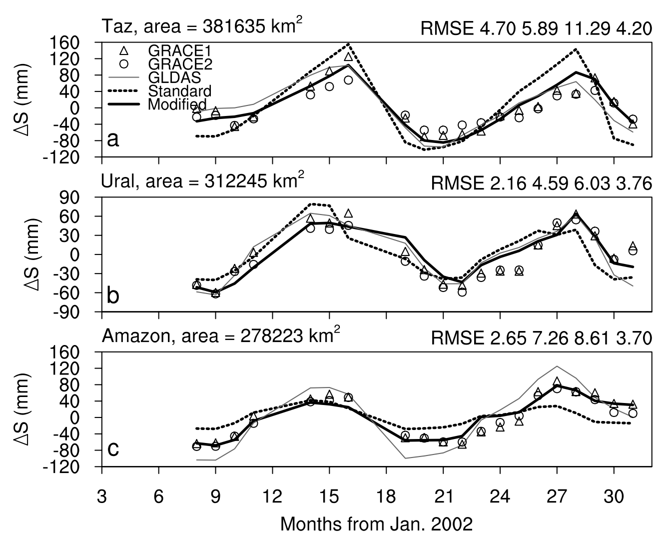

Figure 3. The same as Figure 2, but for three small river basins of Taz, Ural, and Don.

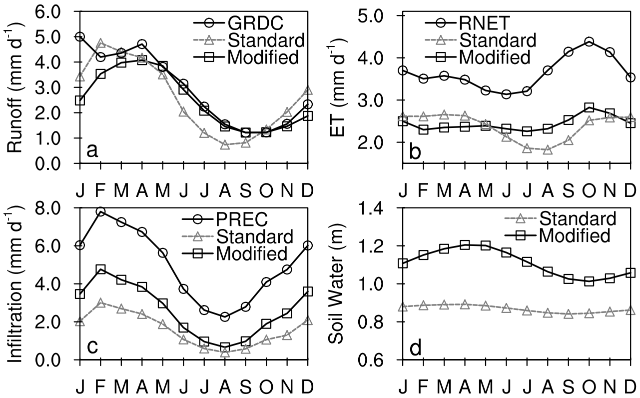

Figure 4. The three-year averaged (a) runoff, (b) *ET*, (c) infiltration, and (d) total soil water in the 3.43 m-deep soil column in the Amazon River modeled with the modified and the standard CLM. The UNH-GRDC runoff climatology (GRDC), net radiation (RNET), and precipitation (PREC) are also included in Figures 4a, 4b, and 4c, respectively.

that from the standard CLM (Figure 4b). This change in *ET* is consistent with the estimated *ET* at a single site (2° 57′ S, 59° 57′ W) that undergoes a weak annual cycle using measured net radiation and atmospheric humidity [*Shuttleworth*, 1988]. The seasonality of the Amazon *ET* is primarily controlled by net radiation (Figure 4b) instead of precipitation (Figure 4c).

[19] The standard CLM produces less soil water and a weaker variability of the soil water than the modified CLM (Figure 4d) due partly to the lower amount of water infiltrated at the soil surface and its weaker variability (Figure 4c). The lower amount of infiltrated water and its weaker variation result from the excessive interception loss of the canopy-intercepted precipitation and the excessive surface runoff [*Niu et al.*, 2005]. Moreover, the extremely large base flow coefficient in the standard CLM produces base flow that further lessens the variability of the soil water, because the base flow is solely proportional to the deep-soil water storage. A sensitivity test using the standard CLM but with (1) a reduced canopy-interception of precipitation by a factor of 5, (2) an enhanced K_{sat} (K_{sat} is defined by the soil texture as in the modified CLM), (3) a decreased surface runoff by a factor of 2, and (4) a reduced base flow coefficient by a factor of 100 (to 4×10^{-4} mm s^{-1}) produces water storage variability similar to that by the modified scheme. Thus, a land surface model that produces a more realistic variability in water storage requires that the hydrological processes, especially the runoff scheme, be better represented and that the hydrologic parameters are better calibrated.

4. Conclusion

[20] We draw conclusions as follows: (1) The modified CLM performs much better in simulating the seasonal variability of TWS in global river basins of various scales, indicating that a model with different representations of runoff and frozen soil may produce fairly different TWS variations. (2) The modified CLM, which simulates more accurate runoff and TWS variations, produces a favorable change in the modeled *ET*, which has a weaker annual cycle than the standard CLM in the Amazon River basin.

[21] The modified CLM does a decent job in simulating TWS variations in the selected river basins although the groundwater is not explicitly represented in the model. However, to what extent the groundwater variation contributes to the TWS variations in regions where the water table is deep is subject to further studies by adding an aquifer model, which describes water storage change in the aquifer, to a land surface model.

[22] **Acknowledgments.** This work was funded by NASA Grant NAG5-10209, NAG5-12577 and NOAA Grant NA03OAR4310076. The authors would like to thank Lindsey E. Gulden for her comments and suggestions. Matthew Rodell is thanked for providing us with the GLDAS observation-derived forcing data. Jianli Chen and K.-W. Seo are thanked for providing us with the GRACE-derived data. The computing resources are provided by the Texas Advanced Computing Center (TACC).

References

Bamzai, A. S., and J. Shukla (1999), Relation between Eurasian snow cover, snow depth, and the Indian summer monsoon: An observational study, *J. Clim.*, *12*(10), 3117–3132.

Chen, J. L., M. Rodell, C. R. Wilson, and J. S. Famiglietti (2005), Low degree spherical harmonic influences on Gravity Recovery and Climate Experiment (GRACE) water storage estimates, *Geophys. Res. Lett.*, *32*, L14405, doi:10.1029/2005GL022964.

Gutowski, W. J., Jr., C. J. Vörösmarty, M. Person, Z. Ötles, B. Fekete, and J. York (2002), A coupled land-atmosphere simulation program (CLASP): Calibration and validation, *J. Geophys. Res.*, *107*(D16), 4283, doi:10.1029/2001JD000392.

Koster, R. D., et al. (2004), Regions of strong coupling between soil moisture and precipitation, *Science*, *305*(5687), 1138–1140.

Niu, G.-Y., and Z.-L. Yang (2006), Effects of frozen soil on snowmelt runoff and soil water storage at a continental scale, *J. Hydrometeorol.*, in press.

Niu, G., Z. Yang, R. E. Dickinson, and L. E. Gulden (2005), A simple TOPMODEL-based runoff parameterization (SIMTOP) for use in global climate models, *J. Geophys. Res.*, *110*, D21106, doi:10.1029/2005JD006111.

Oleson, K. W., et al. (2004), Technical description of the community land model (CLM), *Tech. Note NCAR/TN-461+STR*, 174 pp., Natl. Cent. for Atmos. Res., Boulder, Colo. (Available at www.cgd.ucar.edu/tss/clm/distribution/clm3.0/index.html).

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., et al. (2004b), The global land data assimilation system, *Bull. Am. Meteorol. Soc.*, *85*(3), 381–394.

Seo, K. W., and C. R. Wilson (2005), Simulated estimation of hydrological loads from GRACE, *J. Geod.*, *78*, 442–456.

Shuttleworth, W. J. (1988), Evaporation from Amazonian rainforest, *Proc. R. Soc. London, Ser. B*, *233*, 321–346.

Tapley, B. D., et al. (2004), GRACE measurements of mass variability in the Earth system, *Science*, *305*(5683), 503–505.

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.

G.-Y. Niu and Z.-L. Yang, Department of Geological Sciences, John A. and Katherine G. Jackson School of Geosciences, University of Texas at Austin, Austin, TX 78712–0254, USA. (niu@geo.utexas.edu)