Basin scale estimates of evapotranspiration using GRACE and other observations

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[1] Evapotranspiration is integral to studies of the Earth system, yet it is difficult to measure on regional scales. One estimation technique is a terrestrial water budget, i.e., total precipitation minus the sum of evapotranspiration and net runoff equals the change in water storage. Gravity Recovery and Climate Experiment (GRACE) satellite gravity observations are now enabling closure of this equation by providing the terrestrial water storage change. Equations are presented here for estimating evapotranspiration using observation based information, taking into account the unique nature of GRACE observations. GRACE water storage changes are first substantiated by comparing with results from a land surface model and a combined atmospheric-terrestrial water budget approach. Evapotranspiration is then estimated for 14 time periods over the Mississippi River basin and compared with output from three modeling systems. The GRACE estimates generally lay in the middle of the models and may provide skill in evaluating modeled evapotranspiration. *INDEX TERMS:* 1640 Global Change: Remote sensing; 1818 Hydrology: Evapotranspiration; 1836 Hydrology: Hydrologic budget (1655); 3337 Meteorology and Atmospheric Dynamics: Numerical modeling and data assimilation. **Citation:** Rodell, M., J. S. Famiglietti, J. Chen, S. I. Seneviratne, P. Viterbo, S. Holl, and C. R. Wilson (2004), Basin scale estimates of evapotranspiration using GRACE and other observations, *Geophys. Res. Lett.*, *31*, L20504, doi:10.1029/2004GL020873.

1. Introduction

[2] Evapotranspiration links Earth's water, energy, and carbon cycles. Roughly 50% of the solar radiation incident at the land surface is returned to the atmosphere as latent heat [*Kiehl and Trenberth*, 1997]. Evapotranspiration from the land surface replenishes atmospheric moisture and helps to sustain storms through the process of precipitation recycling [*Brubaker et al.*, 1993; *Eltahir and Bras*, 1996; *Bosilovich and Schubert*, 2002]. It also regulates the spatial-temporal distribution of soil moisture, which is itself a critical lower boundary forcing on climate [*Dirmeyer et al.*, 1999; *Koster et al.*, 2000]. Trends in evapotranspiration rates may be an indicator of climate change, in particular the acceleration of the hydrological cycle and changes in the way heat is redistributed from the tropics to midlatitude and polar regions [*Brutsaert and Parlange*, 1998; *Ohmura and Wild*, 2002; *Roderick and Farquhar*, 2002]. Therefore, improved characterization and quantification of evapotranspiration is essential for improving understanding of Earth system processes.

[3] Nevertheless, evapotranspiration is difficult to estimate at regional (climatic) scales. Micrometeorological measurement networks are generally too sparse for routine monitoring. Remote-sensing approaches typically rely on observations of surface temperature and vegetation indices as input to turbulent transfer or energy balance formulations. Several authors have demonstrated the strengths and weaknesses of these approaches [e.g., *Norman et al.*, 2001; *Kustas et al.*, 2001; *Jiang and Islam*, 2001]. Perhaps the most important limitation is the necessity of region-specific calibration using ancillary data such as air temperature, wind speed, surface resistance parameters, and/or independent estimates of latent and other heat fluxes.

[4] One approach to estimating regional evapotranspiration (ET) is solution of the drainage basin water balance for ET:

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

where P is total precipitation, Q is net stream flow, and ΔS is the change in terrestrial water storage for a specific time period. *Yeh et al.* [1998] applied this method with reasonable success over Illinois, using in situ observations to estimate ΔS. However, while P and Q are often observed with sufficient accuracy to avoid large errors in the residual, independent estimates of ΔS have been lacking for most of the globe.

[5] The Gravity Recovery and Climate Experiment (GRACE) satellites, launched 17 March 2002, are now measuring Earth's gravity field with enough precision to infer terrestrial water mass variations over sufficiently large regions [*Wahr et al.*, 2004; *Tapley et al.*, 2004b]. Groundwater, soil moisture, snow, and surface water all may contribute significantly to the observed ΔS, yet GRACE alone can not separate them, making detailed hydrological interpretation a challenge [*Rodell and Famiglietti*, 2001, 2002]. Whereas ΔS estimates based on modeled or in situ data are apt to overlook one or more components, ΔS derived from GRACE is a perfect fit for water budget studies because it is a horizontally and vertically integrated quantity [*Rodell and Famiglietti*, 1999].

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[6] The objective of this paper is to demonstrate the estimation of ET using a terrestrial water budget approach with observation based ΔS, P, and Q estimates. The Mississippi River basin was chosen as the study region due to the availability of data, however, the method could theoretically be applied to any large drainage basin.

2. Data

[7] Through March 2004, GRACE had delivered 15 near-monthly global gravity field solutions, as sets of Stokes coefficients to a spherical harmonic expansion up to degree and order 120 [e.g., *Tapley et al.*, 2004a]. The effects of atmospheric surface pressure and ocean bottom pressure changes are removed using output from the European Centre for Medium-Range Weather Forecasts (ECMWF) operational forecast model and a barotropic ocean model driven by ECMWF pressure and winds [*Tapley et al.*, 2004a]. The variability in the resulting monthly gravity solutions is introduced mainly by redistribution of terrestrial water mass. The observed gravity signal degrades at higher degrees and orders (shorter length scales), hence there is a tradeoff between signal accuracy and precision in the delineation of a study region. As the selected minimum averaging radius increases, mass changes from outside the region leak into the estimates (leakage error) [e.g., *Swenson et al.*, 2003]. In this study, three averaging radii were considered: 600 km, 800 km, and 1000 km, each representing the half-wavelength of the Gaussian averaging kernel [*Wahr et al.*, 1998]. The degree 2 zonal term Stokes coefficient $C_{2,0}$ was excluded in the computation, because of the relatively large uncertainties associated with this term [*Tapley et al.*, 2004a]. Water storage changes in the Mississippi River basin were directly extracted from the global mass change fields.

[8] The National Oceanic and Atmospheric Administration's (NOAA) Climate Prediction Center's operational global 2.5° 5-day Merged Analysis of Precipitation (CMAP) is the basis for P. This product integrates satellite (infrared and microwave) and gauge observations [*Xie and Arkin*, 1997]. Modeled precipitation fields from NOAA's Global Data Assimilation System (GDAS) [*Derber et al.*, 1991] operational atmospheric analyses were used to disaggregate the CMAP fields to 0.25°, 6-hourly resolutions [*Rodell et al.*, 2004].

[9] Daily discharge measurements for the Mississippi River basin were obtained from the U.S. Army Corps of Engineers (T. Rodgers, personal communication, 2003). These were based on river stage observations from the Vicksburg, Mississippi gauging station.

3. Methods

[10] The change in terrestrial water storage was estimated from GRACE as the difference between one roughly-30-day observation and the previous observation. Accordingly, the monthly basin-scale water balance, neglecting groundwater inflows and outflows, was approximated as

$$\underset{2,1}{S} - \underset{1,1}{S} = \sum_{1,1}^{2,1} P - \sum_{1,1}^{2,1} ET - \sum_{1,1}^{2,1} Q, \qquad (2)$$

where S, P, ET, and Q are daily values, and the first index represents the GRACE observation period and the second the day number of that period. To be exact, the terms on the right side of (2) would be integrals of instantaneous values. At continental drainage basin scales, it can safely be assumed that surface drainage divides coincide with groundwater flow divides, so that inputs and outputs of groundwater can be ignored. Rewriting (2) for all pairs of days in the two observation periods and summing all equations yields

$$\left[\underset{2,1}{S} + \cdots + \underset{2,N}{S}\right] - \left[\underset{1,1}{S} + \cdots + \underset{1,N}{S}\right] = \left[\sum_{1,1}^{2,1} P + \cdots + \sum_{1,N}^{2,N} P\right] - \left[\sum_{1,1}^{2,1} ET + \cdots + \sum_{1,N}^{2,N} ET\right] - \left[\sum_{1,1}^{2,1} Q + \cdots + \sum_{1,N}^{2,N} Q\right], \qquad (3)$$

where N is the number of days per observation period. After dividing both sides by N and simplifying, (4) becomes

$$\Delta\overline{S} = \frac{1}{N} \sum_{n=1}^{N} \sum_{d=D_1+n}^{D_2+n-1} (P_d - ET_d - Q_d), \qquad (4)$$

where $\Delta\overline{S}$ is the change in average water storage, n is the day number of the observation period, d is the date, and D is the first date of the observation period denoted by the index 1 or 2. Equivalently,

$$\Delta\overline{S} = \overline{P} - \overline{ET} - \overline{Q}, \qquad (5)$$

where $\overline{P}$, $\overline{ET}$, and $\overline{Q}$ are running-mean flux accumulations. For example, given two consecutive 30 day periods (60 days), $\overline{ET}$ is the average 30-day evapotranspiration total over all 31 sets of 30 consecutive days within the 60 days. GRACE observation periods are often non-consecutive and different lengths, so that (5) must be expanded to

$$\Delta\overline{S} = \sum_{d=D_1}^{D_1+N_1-1} \frac{d-D_1}{N_1}(P_d - ET_d - Q_d) + \sum_{d=D_1+N_1}^{D_2-1} (P_d - ET_d - Q_d) + \sum_{d=D_2}^{D_2+N_2-1} \frac{D_2+N_2-d}{N_2}(P_d - ET_d - Q_d), \qquad (6)$$

where the indices for D and N denote the observation period. In order to present a daily flux rate, (6) was solved for $\overline{ET}$ (i.e., the sum of all the ET terms) and divided by the effective number of days, $\overline{N}$, contributing to the running mean accumulation, where

$$\overline{N} = \{[N_1 - 1]/2\} + \{[D_2 - (D_1 + N_1)]\} + \{[N_2 + 1]/2\}. \qquad (7)$$

[11] The $\Delta\overline{S}$ term above closely approximates the storage change observed by GRACE, however, the running mean accumulations of (5) are not normally used in hydrology. Therefore, in comparing with output from three numerical models below, mean daily modeled ET rates over the Mississippi River basin for the identical GRACE periods were calculated as equaling the right side of (6) divided by $\overline{N}$, with the P and Q terms set to zero.

[12] Assuming that measurement errors of P, Q, and ΔS are independent and normally distributed about their true values,

$$\sigma_{\overline{ET}}^2 = \sigma_{\overline{P}}^2 + \sigma_{\overline{Q}}^2 + \sigma_{\overline{\Delta S}}^2, \tag{8}$$

where σ is the standard deviation; and

$$\sigma_{\overline{X}} = \frac{\upsilon_{\overline{X}} \overline{X}}{2}, \tag{9}$$

where υ is the relative uncertainty in mean flux accumulation $\overline{X}$. Thus,

$$\upsilon_{\overline{ET}} = \frac{\sqrt{\upsilon_{\overline{P}}^2 \overline{P}^2 + \upsilon_{\overline{Q}}^2 \overline{Q}^2 + \upsilon_{\overline{\Delta S}}^2 \overline{\Delta S}^2}}{\overline{P} - \overline{Q} - \overline{\Delta S}}. \tag{10}$$

Given $\upsilon_{\overline{ET}}$, the relative error for $\overline{ET}$, the 95% confidence limits on $\overline{ET}$ were computed as $\pm \upsilon_{\overline{ET}} \overline{ET}$.

[13] The term $\upsilon_{\Delta \overline{S}} \Delta \overline{S}$ represents the absolute error of a GRACE estimate of monthly water storage change. It includes GRACE instrument and signal retrieval errors, errors from the removal of atmospheric mass variations using model analyses, and leakage errors. *Wahr et al.* [2004] estimated the total error in a single monthly solution to be 1.8 cm for a 750 km averaging radius. This value must be multiplied by $\sqrt{2}$ for the change in mass based on two solutions, hence $\upsilon_{\Delta \overline{S}} \Delta \overline{S}$ was estimated here as 2.5 cm for a 800 km radius. GRACE errors may diminish in the future, as *Tapley et al.* [2004b] showed that refinement of the retrieval technique, which is ongoing, reduced errors significantly between solutions for 2002 and 2003. Uncertainty in the precipitation fields was estimated at 11% (J. Gottschalck et al., unpublished manuscript, 2004). A value of 5% was chosen for uncertainty in the runoff estimates [*Dingman*, 2001].

4. Results

[14] Because ΔS derived from GRACE has not yet been rigorously validated, a comparison with water balance and land surface model based estimates for the Mississippi River basin was performed. A combined atmospheric-terrestrial water balance [e.g., *Rasmusson*, 1968] was used to estimate ΔS with atmospheric convergence fields from the ECMWF operational analysis (see http://www.ecmwf.int/research/ifsdocs/CY25r1/index.html; http://www.ecmwf.int/products/data/operational_system/evolution/index.html) and the runoff data described above as inputs. *Seneviratne et al.* [2004] showed that this approach produced excellent agreement with in situ groundwater, soil moisture, and snow observations averaged over the state of Illinois. A second estimate of ΔS was produced based on soil moisture (2 m column depth), snow, and vegetation canopy surface water output from the Global Land Data Assimilation System (GLDAS) [*Rodell et al.*, 2004] driving the Noah land surface model [*Ek et al.*, 2003].

[15] The resulting time series are shown in Figure 1. In general, the GRACE derived terrestrial water storage anomalies (departures from the series mean) correlate well with the water balance and model based estimates. Based on the fifteen available solutions, the seasonal amplitude of terrestrial water storage inferred from GRACE (800 km averaging radius) lies somewhere between those from the other two approaches. GRACE monthly water storage anomalies based on 1200 km, 1000 km, 800 km, and 600 km averaging radii (not shown) had progressively higher amplitudes at the shorter radii, with a range on the order of 1–2 cm. This suggests that leakage errors have the effect of reducing the amplitude of the inferred anomalies, so that the true seasonal amplitude may be larger than what was estimated here. Nevertheless, it is safe to conclude that GRACE does measure ΔS with some acceptable level of error.

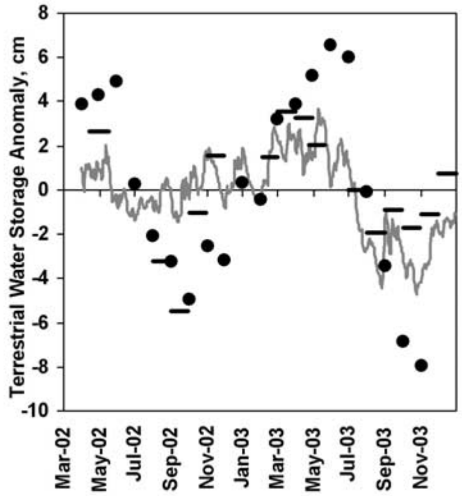

Figure 1. Time series of terrestrial water storage anomalies (departures from the series means) from the GRACE water balance approach (horizontal lines), combined atmospheric-terrestrial water balance approach (dots), and GLDAS/Noah land surface model (gray line).

[16] Estimates of mean daily ET based on the GRACE water balance approach are plotted in Figure 2, along with estimates from GLDAS/Noah, ECMWF, and GDAS atmospheric analyses [*Derber et al.*, 1991]. The horizontal bars at the bottom of the plot indicate the time period contributing to each estimate, and the Y error bars represent the 95% confidence interval. The GRACE estimates of ET generally fall in the middle of the model estimates. The RMS differences between GRACE and GLDAS/Noah, GDAS, and ECMWF are 0.83, 0.67, and 0.65 mm/day, while the standard deviation of the three models averages 0.63 mm/day. The uncertainty in the GRACE estimates is on the same order, ranging from 0.33 to 1.33 mm/day and averaging 0.86 mm/day. ECMWF shows the best agreement with GRACE and tends to underestimate ET, with a mean bias of -0.43 mm/day. GLDAS/Noah also underestimates ET, with a mean bias of -0.62 mm/day, while GDAS has a positive bias of 0.55 mm/day. Estimating ET using the 600 km, 1000 km, and 1200 km averaging radii for GRACE (not shown) had little effect on the resulting model RMS differences and biases. Hence this technique

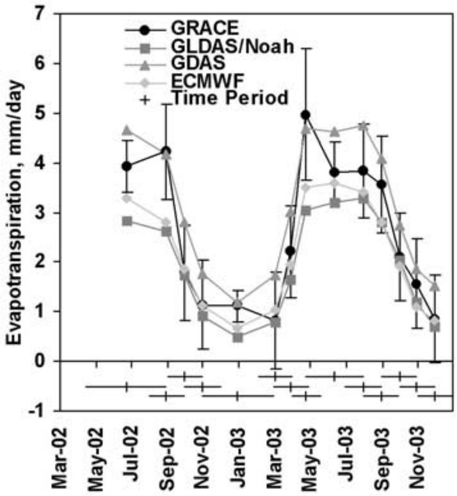

Figure 2. GRACE and model based estimates of evapotranspiration rate. The horizontal bars at the bottom indicate the time periods contributing to the averages, and the Y error bars indicate the uncertainty in the GRACE estimates.

is able to provide independent estimates of regional *ET* which are likely to be valuable as a validation tool.

5. Summary

[17] Equations were developed for estimating evapotranspiration (*ET*) using a water balance approach with terrestrial water storage changes derived from GRACE and observation based precipitation and runoff. GRACE water storage change estimates were shown to compare favorably with results from a land surface model and a combined atmospheric-terrestrial water balance approach. *ET* was estimated for 14 time periods over the Mississippi River basin, and generally lay in the middle of estimates from a land surface model and two operational atmospheric analysis systems. Uncertainty in GRACE based *ET* was estimated to be 0.86 mm/day. Biases in the model estimates were consistent over time and on the same order as the GRACE uncertainty, so that it was concluded that the technique is valuable for assessing modeled *ET*.

References

Bosilovich, M. G., and S. D. Schubert (2002), Water vapor tracers as diagnostics of the regional hydrologic cycle, *J. Hydrometeorol.*, *3*, 149–165.

Brubaker, K. L., D. Entekhabi, and P. S. Eagleson (1993), Estimation of continental precipitation recycling, *J. Clim.*, *6*, 1077–1089.

Brutsaert, W., and M. B. Parlange (1998), Hydrologic cycle explains the evaporation paradox, *Nature*, *396*, doi:10.1038/23845.

Derber, J. C., D. F. Parrish, and S. J. Lord (1991), The new global operational analysis system at the National Meteorological Center, *Weather Forecasting*, *6*, 538–547.

Dingman, S. L. (2001), *Physical Hydrology*, Prentice-Hall, Englewood Cliffs, N. J.

Dirmeyer, P. A., A. J. Dolman, and N. Sato (1999), The pilot phase of the global soil wetness project, *Bull. Am. Meteorol. Soc.*, *80*, 851–878.

Ek, M. B., K. E. Mitchell, Y. Lin et al. (2003), Implementation of the upgraded Noah land-surface model in the NCEP operational mesoscale Eta model, *J. Geophys. Res.*, *108*(D22), 8851, doi:10.1029/2002JD003296.

Eltahir, E. A. B., and R. L. Bras (1996), Precipitation recycling, *Rev. Geophys.*, *34*, 367–378.

Jiang, L., and S. Islam (2001), Estimation of surface evaporation map over southern great plains using remote sensing data, *Water Resour. Res.*, *37*, 329–340.

Kiehl, J. T., and K. E. Trenberth (1997), Earth's annual global mean energy budget, *Bull. Am. Meteorol. Soc.*, *78*, 197–208.

Koster, R. D., M. J. Suarez, and M. Heiser (2000), Variance and predictability of precipitation at seasonal-to-interannual timescales, *J. Hydrometeorol.*, *1*, 26–46.

Kustas, W. P., G. R. Diak, and J. M. Norman (2001), Time difference methods for monitoring regional-scale heat fluxes with remote sensing, in *Land Surface Hydrology, Meteorology and Climate: Observations and Modeling*, *Water Sci. Appl.*, vol. 3, edited by V. Lakshmi, J. Albertson, and J. Schaake, pp. 15–30, AGU, Washington, D. C.

Norman, J. M., W. P. Kustas, J. H. Prueger, and G. R. Diak (2001), Surface flux estimation using radiometric temperature: A dual-temperature-difference method to minimize measurement errors, *Water Resour. Res.*, *36*, 2263–2274.

Ohmura, A., and M. Wild (2002), Is the hydrological cycle accelerating?, *Science*, *298*, 1345–1356.

Rasmusson, E. M. (1968), Atmospheric water vapor transport and the water balance of North America. II. Large-scale water balance investigations, *Mon. Weather Rev.*, *96*, 720–734.

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*, 2705–2723.

Rodell, M., and J. S. Famiglietti (2001), An analysis of terrestrial water storage variations in Illinois with implications for the Gravity Recovery and Climate Experiment (GRACE), *Water Resour. Res.*, *37*, 1327–1340.

Rodell, M., and J. S. Famiglietti (2002), The potential for satellite-based monitoring of groundwater storage changes using GRACE: The High Plains aquifer, central U.S., *J. Hydrol.*, *263*, 245–256.

Rodell, M., et al. (2004), The Global Land Data Assimilation System, *Bull. Am. Meteorol. Soc.*, *85*, 381–394.

Roderick, M. L., and G. D. Farquhar (2002), The cause of decreased pan evaporation over the past 50 years, *Science*, *298*, 1410–1412.

Seneviratne, S., P. Viterbo, D. Lüthi, and C. Schär (2004), Inferring changes to terrestrial water storage using ERA-40 reanalysis data: The Mississippi River basin, *J. Clim.*, *17*, 2039–2057.

Swenson, S., J. Wahr, and P. C. D. Milly (2003), Estimated accuracies of regional water storage variations inferred from the Gravity Recovery and Climate Experiment (GRACE), *Water Resour. Res.*, *39*(8), 1223, doi:10.1029/2002WR001808.

Tapley, B. D., S. Bettadpur, M. Watkins, and C. Reigber (2004a), The Gravity Recovery and Climate Experiment: Mission overview and early results, *Geophys. Res. Lett.*, *31*, L09607, doi:10.1029/2004GL019920.

Tapley, B. D., S. Bettadpur, J. C. Ries et al. (2004b), GRACE measurements of mass variability in the Earth system, *Science*, *305*, 503–505.

Wahr, J., M. Molenaar, and F. Bryan (1998), Time-variability of the Earth's gravity field: Hydrological and oceanic effects and their possible detection using GRACE, *J. Geophys. Res.*, *103*, 30,205–30,230.

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.

Xie, P., and P. A. Arkin (1997), Global precipitation: A 17-year monthly analysis based on gauge observations, satellite estimates and numerical model outputs, *Bull. Am. Meteorol. Soc.*, *78*, 2539–2558.

Yeh, P. J. F., M. Irizarry, and E. A. B. Eltahir (1998), Hydroclimatology of Illinois: A comparison of monthly evaporation estimates based on atmospheric water balance and soil water balance, *J. Geophys. Res.*, *103*, 19,823–19,837.

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