

1           **Water balance of the Arctic drainage system using GRACE**  
2   **gravimetry products**

3  
4   FREDERIC FRAPPART \*†, GUILLAUME RAMILLIEN ‡, JAMES S. FAMIGLIETTI †

5  
6   Submitted to International Journal of Remote Sensing, October, 2009

7  
8   Affiliations :

9  
10 †Department of Earth System Science, University of California, Irvine, Croul Hall  
11 Irvine, CA 92697-3100, USA.

12  
13 ‡Dynamique Terrestre et Planétaire (DTP), Observatoire Midi-Pyrénées, UMR 5562,  
14 CNES/CNRS/IRD/UPS, 14 Av. Edouard Belin, 31400 Toulouse, France.

15  
16 (\*) Now at Laboratoire des Mécanismes et Transferts en Géologie (LMTG), UMR 5563,  
17 CNRS/IRD/UPS, 14 Av. Edouard Belin, 31400 Toulouse, France.

18  
19  
20  
21  
22  
23  
24  
25   Corresponding author : [frederic.frappart@lmtg.obs-mip.fr](mailto:frederic.frappart@lmtg.obs-mip.fr)

26 **Abstract:** Land water and snow mass anomalies versus time were computed from the  
27 inversion of 50 GRACE geoids (08/2002 - 02/2007) from the RL04 GFZ release and used  
28 to characterize the hydrology of the Arctic drainage system. GRACE-based time series  
29 have been compared to Snow Water Equivalent and snow depth climatologies, and  
30 snowfall for validation purpose. Time series of regional averages of water volume were  
31 estimated for the eleven largest Peri-Arctic basins. Strong correlations were found  
32 between the snow estimates and river discharges in the Arctic basins (0.49 to 0.8). Then  
33 changes in land waters storage have been compared to precipitation minus  
34 evapotranspiration fluxes to determine which flux of the hydrological budget controls the  
35 Arctic hydrology. Results are very contrasted according to the basin. Trends of snow and  
36 land water masses were also computed over the 2003-2006 period. Eurasian basins loose  
37 snow mass whereas North American basins are gaining mass.  
38

39

## 1. Introduction

The Arctic region is a major component of the global climate system and is expected to be importantly affected by global warming (Peterson *et al.*, 2002). Although the Arctic Ocean holds only 1% of global volume of seawater, it receives 11% of the world's freshwater input (Lammers *et al.*, 2001). The Arctic rivers discharges contribute 50% to the net flux of freshwater into the Arctic Ocean (Barry and Serreze, 2000). Arctic hydrological systems exhibit large temporal variability caused by large-scale changes in atmospheric circulation (Proshutinsky *et al.*, 1999). Discharge observations indicate a significant increase in Arctic discharge since the mid-1930's, with an acceleration in the recent decades (Peterson *et al.*, 2002; Serreze *et al.*, 2003; McClelland *et al.*, 2004; Stocker and Raible, 2005). Timing and magnitude of northern river streamflow are mostly influenced by winter snow mass storage and its subsequent melt (Rango, 1997; Cao *et al.*, 2002; Yang *et al.*, 2003; 2007; Déry *et al.*, 2005; Dyer, 2008; Yang *et al.*, 2009). The snow melt and associated floods during the spring/summer period are the most important hydrologic event of the year in the northern river basins (Cao *et al.*, 2002; Yang *et al.*, 2003). Changes in pattern of snow cover at high latitudes, such as the earlier start of snowmelt associated with warming in winter and spring seasons (Lammers *et al.*, 2001; Kitaev *et al.*, 2005; Groisman *et al.*, 2006; Bulygina *et al.*, 2007), may accentuate the variability of hydrologic regime at high latitudes in the context of global warming (Barnett *et al.*, 2005).

The launch of the Gravity Recovery and Climate Experiment (GRACE) space mission in March 2002 enables, for the first time, detection of tiny temporal variations in Earth's gravity field (Tapley *et al.*, 2004 a, b), which over land are mainly due to the vertically-integrated water mass changes inside aquifers, soil, surface reservoirs and snow pack, if effects of noise and residual errors from correcting models for atmosphere and ocean masses are neglected (Wahr *et al.*, 1998; Rodell and Famiglietti, 1999; Swenson *et al.*, 2003). Wahr *et al.*, 2006 have shown that GRACE data over the continents provide information on the total land water storage with an accuracy of 15 - 20 mm of water thickness equivalent a spatial Gaussian average with a radius of 400 km. Ramillien *et al.* (2005) used an iterative inverse approach to estimate variations in continental water storage (i.e. all of the groundwater, soil water, surface water, snow and ice) and separate land waters and snow components from the GRACE RL02 data. Comparisons with model outputs and microwave observations have already demonstrated the quality of RL03 and RL02 land water and snow solutions derived by inverse method (Ramillien *et al.* 2005, 2006; Frappart *et al.*, 2006). In a recent study, Niu *et al.* (2007) showed that the spatial pattern of snow derived from GRACE has a better agreement with climatologies than passive microwave estimates.

Our goal is to study the consistency of the snow mass variations derived from GRACE in terms spatial and temporal patterns. In this study, we will be able, for the first time, to compare direct measurements of total land water and snow storages with river discharges in the Arctic drainage system. Previously, Syed *et al.* (2007) estimated river discharge from several Arctic basins, and compared GRACE-derived land water storage (but not separate out snow storage) to observed and estimated discharge. In the present work we more directly characterize the relationship between total land water, snow storage and river discharge. We use the RL04 GRACE land water and snow solutions

86 computed using the method developed by Ramillien *et al.* (2005, 2006) to estimate time  
 87 series of basin-scale land water and snow volume anomalies. We present estimates of  
 88 Snow Water Equivalent (SWE) and Terrestrial Water Storage (TWS) anomalies from  
 89 August 2002 to February 2007 for the eleven largest Arctic drainage basins, i.e., Yukon,  
 90 Mackenzie, Nelson, Severnyy Dvina, Pechora, Ob, Yenisey, Kotya, Lena, Indigirka,  
 91 Kolyma (figure 1). We validated the GRACE-derived snow solutions by comparing them  
 92 with pan-Arctic snow depth climatologies from USAF/ETAC and the Arctic Climatology  
 93 Project, a SWE climatology over North America and snowfall. While previous work has  
 94 focused on the relationship between snow extent or depth and river runoff (Yang *et al.*,  
 95 2003; Déry *et al.*, 2005; Grippa *et al.*, 2005), we compare continental water storage and  
 96 snow volume variations derived from the inversion of GRACE geoids to *in situ* discharge  
 97 for the largest Arctic river basins.

## 98 2. Datasets

99

### 100 2.1 GRACE-derived land water and snow mass solutions

101 We use the [monthly](#) land water and snow solutions derived from the inversion of 50  
 102 GRACE geoids from the fourth data release by GeoForschungZentrum (GFZ-RL04), as  
 103 presented in Ramillien *et al.* (2005, 2006). These solutions range from August 2002 to  
 104 February 2007, with a few missing months (September and December 2002, January,  
 105 June and July 2003, January 2004). [They represent anomaly of mass expressed in terms](#)  
 106 [of equivalent water thickness.](#)

107 The GRACE-based land water and snow solutions separately computed in  
 108 Ramillien *et al.* (2005) are spherical harmonics of a surface density function  $F(\theta, \lambda, k)$   
 109 that represents the global map of either land waters or snow mass:

$$110 \quad F(\theta, \lambda, k) = \sum_{n=1}^N \sum_{m=0}^n [C_{nm}^F(k) \cos(m\lambda) + S_{nm}^F(k) \sin(m\lambda)] \tilde{P}_{nm}(\cos \theta) \quad (1)$$

111 In Equation 1,  $\theta$  and  $\lambda$  are co-latitude and longitude,  $k$  is the number of a given monthly  
 112 solution.  $n$  and  $m$  are degree and order,  $\tilde{P}_{nm}$  is the associated Legendre function, and  
 113  $C_{nm}^F(t)$  and  $S_{nm}^F(t)$  are the normalized water (or snow) mass coefficients (units: mm of  
 114 equivalent water height) which were estimated by inversion (Ramillien *et al.*, 2005). In  
 115 practice, the spherical harmonic development cutoff  $N$  used for the land water solutions is  
 116 limited to degree  $N=50$ . This corresponds to a spatial resolution of  $\sim 400$  km at the surface  
 117 of the Earth. The GRACE-based land water and snow maps were interpolated on  $1^\circ \times 1^\circ$   
 118 regular grids.

119

### 120 2.2 Snow depth climatologies

#### 121 2.2.1 Global snow depth multi-year average

122

123 USAF/ETAC (United States Air Force/Environmental Technical Applications  
 124 Center USAF/ETAC) climatology is a  $1^\circ \times 1^\circ$  monthly gridded dataset composed of snow  
 125 depths averaged over an approximately 30-year window ending in the 1980s. The data

126 comes from various sources with varying degrees of accuracy, and was manually edited  
127 and interpolated using relatively simple methods (Foster and Davy, 1988).

### 128 **2.2.2 American-Russian snow depth climatology**

129 The Environmental Working Group (EWG) Climatology Project compiled data  
130 on Arctic regions to expand scientific understanding of the Arctic and edited a set of  
131 complementary atlases for Arctic oceanography, sea-ice, and meteorology, under the  
132 framework of the U.S.-Russian Joint Commission on Economics and Technological  
133 Cooperation (Arctic Climatology Project, 2000). The snow climatology is a gridded  
134 dataset in ASCII EASE Grid format with a cell size of 250 km of monthly mean snow  
135 depth fields over the period 1966-1982.

### 136 **2.2.3 Gridded Monthly SWE climatology over North America**

137  
138 The Canadian Meteorological Service developed an operational snow depth  
139 analysis scheme which uses extensive daily snow depth observations from Canada and  
140 the USA to generate grids of snow depths and SWE at a resolution of 0.25° (Brassnet,  
141 1999). The monthly climatology grids were derived from daily snow depth and SWE  
142 grids covering the hydrological years 1979/80 to 1996/97. The gridded output is  
143 dominated by observations South of about 55° N. North of 55° N, the output is dominated  
144 by the snow model. SWE was estimated using the density values simulated by the snow  
145 model (Brown *et al.*, 2003).

### 146 **2.3 Snowfall derived from GPCP rainfall**

147  
148 The Global Precipitation Climatology Project (GPCP), established in 1986 by the  
149 World Climate Research Program, provides data that quantify the distribution of  
150 precipitation over the whole globe (Adler *et al.*, 2003). We use here the Satellite-Gauge  
151 Combined Precipitation Data product of GPCP Version 2 data for evaluating our  
152 estimates of monthly SWE variations in the pan-Arctic region. The GPCP products we  
153 are using are monthly means with a spatial resolution of 1° of latitude and longitude and  
154 are available from January 1979 to present. Over land surfaces, the uncertainty in the rate  
155 estimates from GPCP is generally lower than over the oceans due to the *in situ* gauge  
156 input (in addition to satellite) from the GPCC (Global Precipitation Climatology Center).  
157 Over land, validation experiments have been conducted in a variety of location  
158 worldwide and suggest that while there are known problems in regions of persistent  
159 convective precipitation, non precipitating cirrus or regions of complex terrain, the  
160 estimates uncertainties range between 10%–30% (Adler *et al.*, 2003).

161 Monthly snowfall is estimated from GPCP rainfall using the NCEP air  
162 temperature topographically adjusted (available from the Arctic Rims website:  
163 <http://rims.unh.edu/>) according to the following equation:

164

$$P_{snow} = P_{tot} \theta(T_0 - T) \text{ with } \theta = \begin{cases} 0 & \text{if } T > T_0 = 2^\circ\text{C} \\ \frac{T_0 - T}{2} & \text{if } 0 < T < T_0 \\ 1 & \text{if } T > 0 \end{cases} \quad (2)$$

165 where  $P_{snow}$  is the estimated snowfall,  $P_{tot}$  is the GPCP rainfall,  $\theta$  is a threshold function  
 166 of air temperature,  $T$  the air temperature and  $T_0$  the threshold air temperature ( $0^\circ\text{C}$ ).

167 **2.4 Snow outputs from WGHM model**

168

169 The Water GAP Global Hydrology Model (WGHM) computes  $0.5^\circ \times 0.5^\circ$   
 170 gridded time series of monthly runoff and river discharge and is tuned against time series  
 171 of annual river discharges measured at 724 globally-distributed stations (Döll *et al.*,  
 172 2003). It also provides monthly grids of snow and soil water. The effect of snow is  
 173 simulated by a simple degree-day algorithm. Below  $0^\circ\text{C}$ , precipitation falls as snow and  
 174 is added to snow storage. Above  $0^\circ\text{C}$ , snow melts with a rate of 2 mm/day per degree in  
 175 forests and of 4 mm/day in case of other land cover types. These monthly gridded data  
 176 are available from January 2002 to June 2006.

177 **2.5 River discharge measurements**

178

179 The monthly river discharge measurements at the closest station to the mouth of  
 180 each basin were obtained at the Arctic RIMS (Rapid Integrated Monitoring System)  
 181 website (ArcticRIMS, 2003) for the eleven largest Peri-Arctic drainage basins which has  
 182 developed a near-real time monitoring of pan- Arctic water budgets and river discharge to  
 183 the Arctic Ocean. The availability of the data for each basin is reported in table 1.

184 **2.6 Precipitation minus evapotranspiration dataset**

185

186 This dataset provides estimates of monthly precipitation minus evapotranspiration  
 187 (P-ET) parameter using wind and humidity data from the NCEP/NCAR reanalysis with  
 188 the "aerological method" developed by Kalnay *et al.* (1996). The P-ET parameter is  
 189 equivalent to the vertically-integrated vapor flux convergence adjusted by the time  
 190 change in precipitable water. On monthly timescales, P-ET is dominated by the flux  
 191 convergence term. NCEP/NCAR archives of vertical integrals of the monthly-mean zonal  
 192 and meridional fluxes and precipitable water (based on 6-hourly values at sigma levels),  
 193 are used to compute the flux differences. The P-ET fields are interpolated to the 25 km  
 194 EASE grid. Details of the P-ET calculations and some climate applications are provided  
 195 by Cullather *et al.* (2000) and Serreze *et al.* (2003).

196 **2.7 Post-glacial rebound model**

197

198 The Post-Glacial Rebound (PGR) designates the rise of land masses that were  
 199 depressed by the huge weight of ice sheets during the last glacial period that ended  
 200 between 10,000 and 15,000 years ago. It corresponds to a vertical elevation of the crust

201 which happens especially in Scandinavia, the Hudson Bay in Canada (and maybe  
202 Antarctica) and affects the long wavelength components of the gravity field.

203 The PGR model used in this study is made available by the GRACE Tellus  
204 website (<http://grace.jpl.nasa.gov>). This model is based on Paulson *et al.* (2007) study  
205 and uses the global ICE-5G deglaciation model of Peltier (2004). It assumes an  
206 incompressible, self-gravitating Earth. The mantle is a Maxwell solid, and overlies an  
207 inviscid core. More details on ICE-5G can be found in Peltier (2004). Effects of a  
208 dynamic ocean response through the sea level equation were included using the  
209 formulation of polar wander described by Mitrovica *et al.* (2005). Uncertainty on its  
210 estimates is supposed to be around 20% (Paulson *et al.*, 2007).

211 The GRACE Tellus website provides estimates of the rate of change of surface  
212 mass, expressed in  $\text{mm.yr}^{-1}$  of equivalent water thickness. Degree-one terms were omitted  
213 when computing the mass, because they are not included in the GRACE solutions. The  
214 results were smoothed using a Gaussian averaging function of 500 km radius. The mass  
215 estimates are provided on a 1 x 1 degree grid, spaced a half-degree apart.

### 216 3. Validation of the GRACE-based Snow Water Equivalent

#### 217 3.1 Annual cycle of GRACE-based SWE and comparisons with climatologies 218 and GPCP-derived snowfall

219 From the series of SWE anomaly grids derived from GRACE (using equation 1),  
220 the temporal trend, seasonal and semi-annual amplitude were simultaneously fitted by  
221 least-square adjustment at each grid point. We assumed that, at 1<sup>st</sup> order, the changes of  
222 SWE  $\delta q(t)$  at each grid point are the sum of a linear trend, an annual sinusoid (which  
223 pulsation is  $\omega_{ann} = \frac{2\pi}{T_{ann}}$ , with  $T_{ann} \sim 1$  year), a semi-annual sinusoid (which pulsation is

224  $\omega_{semi-ann} = \frac{2\pi}{T_{semi-ann}}$ , with  $T_{semi-ann} \sim 6$  months) and water mass residuals  $\delta q^{RES}(t)$  :

$$225 \delta q(t) = At + B + C \cos(\omega_{ann} t + \varphi_{ann}) + D \cos(\omega_{semi-ann} t + \varphi_{semi-ann}) + \delta q^{RES} \quad (3)$$

226 The parameters which we adjusted for each grid point ( $\theta, \lambda$ ) are the linear trend  
227 (i.e. slope  $A$  and y-intercept  $B$ ), the annual cycle (i.e. amplitude  $C$  and phase  $\varphi_{ann}$ ) and the  
228 semi-annual cycle (i.e. amplitude  $D$  and phase  $\varphi_{semi-ann}$ ). For this purpose, we used a least-  
229 squares fitting to solve the system:

$$230 \delta Q = \Phi \cdot X \quad (4)$$

231 where the vector  $\delta Q$  is the list of the SWE values,  $\Phi$  and  $X$  are the configuration matrix  
232 and the parameter vector, respectively. The latter two terms are:

$$233 \{\Phi_j\} = \begin{bmatrix} t_j & 1 & \cos(\omega_{ann} t_j) & \sin(\omega_{ann} t_j) & \cos(\omega_{semi-ann} t_j) & \sin(\omega_{semi-ann} t_j) \end{bmatrix} \quad (5a)$$

$$234 X = \begin{bmatrix} \alpha & \beta & \xi_{ann} \cos \varphi_{ann} & -\xi_{ann} \sin \varphi_{ann} & \xi_{semi-ann} \cos \varphi_{semi-ann} & -\xi_{semi-ann} \sin \varphi_{semi-ann} \end{bmatrix} \quad (5b)$$

235 for adjusting the temporal trend and for fitting the annual and semi-annual amplitude and  
236 phase.

237 According to the least-squares criteria, the solution vector of the system is:

$$X^{SOL} = (\Phi^T \Phi)^{-1} \Phi^T \delta Q \quad (6)$$

To locate the regions of snow accumulation, we focused on the annual cycle of SWE at high latitudes. Figure 2a presents the map of amplitude of annual cycle of SWE derived from the inversion of 4 years (2003-2006) of GRACE geoids. The two largest maxima of annual amplitude (~100 mm) are located over North America in the northern part of the Rocky Mountains and the Western part of Canada. Over Eurasia, maximal amplitudes (70-90 mm) are observed in the eastern part of the Ob, Yenisey basins and the Kolyma basins. Secondary maxima, reaching 60 mm of SWE, are present in the Western part of the Eurasian continent (Scandinavia, Severnyy Dvina, Pechora and the Western part of the Ob basins).

Due to the coarse spatial and temporal resolutions (respectively 400 km and one month) of the GRACE-derived snow mass estimates, an indirect validation have been made using climatologies of snow depth from USAF/ETAC and EWG and snowfall-derived from GPCP rainfall products over North America and Eurasia, and a climatology of SWE over North America.

Figure 2 also presents the mean map of annual of snow depth from USAF/ETAC (b) and EWG (c) climatologies, and the total annual snowfall derived from GPCP rainfall over the 2003-2006 period (d). The characteristics of these datasets are summarized in table 2. For comparison purpose, all the datasets have been resampled to a spatial resolution of 1°. The amplitude of annual cycles of GRACE-derived SWE, snow depth from both climatologies and snowfall derived from GPCP show similar patterns. The linear correlation coefficients between the GRACE amplitude of annual cycle and the mean annual snow depths from USAF/ETAC, EWG, and the total snowfall derived from GPCP are respectively 0.53, 0.42, and 0.37. A strong signal can be observed on Eastern Canada (Newfound Land, Labrador and Baffin Island), Scandinavia, river basins in the European part of Russia (Severnyy Dvina and Pechora) and the Yenisey basin on all the datasets. On the contrary, locations of snow accumulation are quite different between GRACE-derived SWE and snow depth climatologies on North-West Canada and East Siberia. Over North America, snow depth climatologies present a strong signal on Alaska and Yukon, Mackenzie and Nelson basins whereas GRACE-derived SWE has a strong maximum on the Rocky Mountains. Over Eurasia, USAF/ETAC snow climatology presents large snow depths on East Siberia (Kotia, Lena, Indigirka and Kolyma basins), EWG has the same pattern except for Indigirka basin, whereas GRACE-derived SWE presents lower snow accumulations over these regions. Two factors can explain these differences: the time periods considered (1950-1980 for USAF/ETAC climatology, 1966-1982 for EWG climatology, and 2003-2006 for both GRACE-derived SWE and GPCP-derived snowfall) in regions which have a strong response to climate variability and the quantities compared related by the snow density which exhibits strong variability both in space and time.

Figure 3 displays the timing in months when occurs the maximum of the GRACE-derived SWE and snow depth from climatologies. Similar patterns can be observed on the three maps, especially a North-South gradient with maximum of snow occurring later in the North than in the South. The major difference lies in maxima occurring sooner in most of Siberia, Alaska and the North of the Rocky Mountains in the GRACE-derived SWE than in the snow depth climatologies. This is in accordance with



286 the decrease of snow cover observed over Siberia between 1956 and 2004. This decrease  
 287 was especially strong over central Siberia in late spring (April-May) for the period 1956-  
 288 1991 (Groisman *et al.*, 2006).

289 A comparison between GRACE-derived SWE and a monthly mean climatology  
 290 over North America was achieved. Figure 4 exhibits the amplitude of annual cycle of  
 291 GRACE-derived SWE and the mean annual of the SWE climatology. Both spatial  
 292 pattern and intensity are very similar between the two products with a correlation  
 293 coefficient of 0.58. The major difference is the strong signal over Alaska present in the  
 294 climatology and lacking in the GRACE product. This difference can be explained by the  
 295 sparse coverage of stations in this region (Brown *et al.*, 2003) and the time period  
 296 considered as an increase of 0.4°C for the mean winter temperature has been observed  
 297 between 1977 and 2004 (Molnia *et al.*, 2007) which caused a decrease in the depth of the  
 298 snow cover in Alaska (Osterkamp, 2005).  
 299

### 300 3.2 Basin scale SWE time-series

301  
 302 For a given month  $t$ , regional average of land water or snow volume (or height)  
 303  $\delta V(t)$  ( $\delta h(t)$  respectively) over a given river basin of area  $A$  is simply computed from the  
 304 water height  $\delta h_j$ , with  $j=1, 2, \dots$  (expressed in terms of mm of equivalent-water height)  
 305 inside  $A$ , and the elementary surface  $R_e^2 \delta\lambda \delta\theta \sin\theta_j$  :

$$306 \quad \delta V(t) = R_e^2 \sum_{j \in A} \delta h_j(\theta, \lambda, t) \sin\theta_j \delta\lambda \delta\theta \quad (7a)$$

$$307 \quad \delta h(t) = \frac{R_e^2}{A} \sum_{j \in A} \delta h_j(\theta, \lambda, t) \sin\theta_j \delta\lambda \delta\theta \quad (7b)$$

308 where  $\theta$  and  $\lambda$  are co-latitude and longitude,  $\delta\lambda$  and  $\delta\theta$  are the grid steps in longitude and  
 309 latitude respectively (generally  $\delta\lambda = \delta\theta$ ). In practice, all points of  $A$  used in (equation 7a  
 310 and equation 7b) are extracted for the eleven drainage basins masks at a 0.5° resolution  
 311 provided by Oki and Sud (1998).

312 Figure 5 presents GRACE-based SWE time series for the 4 largest Arctic  
 313 river basins (Ob, Yenisey, Lena, Mackenzie). In view of the short time span considered  
 314 here, the signal is dominated by the seasonal component with maxima of snow observed  
 315 in February or March for all the basins. We estimated correlation between the time series  
 316 of GRACE-derived SWE and the time series of GPCP-derived snowfall for each basin,  
 317 using the cross-correlation function:

$$318 \quad \Gamma(t) = (S * D)(t) = \int_{\Delta t} S^*(\tau) D(\tau - t) d\tau \quad (8)$$

319 where  $\Gamma$  is the cross-correlation of the SWE  $S$  and the snowfall  $D$  at month  $t$ ,  $\tau$  is the time  
 320 and  $\Delta t$  is the considered period of integration.

321 The time lag between snow and discharge peaks corresponds to the month  $t_0$  that  
 322 maximizes the cross-correlation function  $\Gamma$ :

$$323 \quad \Gamma_{\max} = \Gamma(t_0) = \max_{t \in \Delta t} (\Gamma(t)) \quad (9)$$

324 where  $\Gamma_{\max}$  is the maximum of  $\Gamma$  over the period  $\Delta t$ .

325 The results obtained are presented in table 3 for the maximum of correlation and  
 326 the time lag between the peak of snowfall and the peak of SWE. For most of the basins,  
 327 an agreement better than 40% is generally observed between snowfall and SWE (except

328 for the Nelson basin). The time lags between snowfall and SWE never exceed 2 months.  
329 Due to the spatial resolution of GRACE products, the bigger is the basin, the higher is the  
330 correlation (greater than 0.6, except for Lena). The exceptions are the Nelson and  
331 Indigirka basins where little or no SWE winter peak is observed in the GRACE-derived  
332 product. We compared the SWE derived from GRACE measurements with the SWE  
333 estimated by the WGHM model used as initial guess in the inverse method to extract the  
334 different hydrological components from GRACE data. For the Nelson and Indigirka  
335 basins, the amplitude of the snow signal from WGHM is also low (figure 6). Wahr *et al.*  
336 (2006) estimated that the accuracy of GRACE geoids is 15-20 mm water equivalent  
337 height for a spatial Gaussian average with a radius of 400 km. In these two cases, the  
338 monthly amplitude of the SWE signal is most of the time lower than 20 mm which  
339 represents the limit of detectability of a hydrological signal in the GRACE products.

#### 340 **4. Analysis of water storage changes in the Arctic drainage system**

##### 341 ***4.1 Basin scale TWS, snow and river discharge time series***

342  
343 Figure 7 compares the monthly time series of SWE anomalies and TWS  
344 anomalies derived from GRACE measurements with the total water volume transferred to  
345 the Arctic Ocean for six Arctic drainage basins where river discharges measurements are  
346 available, i.e., Ob, Yenisey, Lena, Mackenzie, Severnaia Dvina and Kolyma. The total  
347 volume of water that flows from a basin to the Arctic Ocean each month is simply  
348 computed as the time integrated river discharge during the month.

349 Maxima of snow are observed in February or March for all the basins whereas  
350 maxima of land waters occurred generally one month later. These results tallied with  
351 those obtained by Dyer *et al.*, 2008 over the Yukon and Mackenzie basin with maximum  
352 of snow depth respectively occurring around day  $(55 \pm 25)$  and  $(65 \pm 15)$  over the period  
353 1975-2000 and 1972-2000. They are also in accordance with peak of SWE estimated  
354 using passive microwave observations for the Yukon (week 8 to 12), Ob (week 8), Lena  
355 and Yenisey (week 7) basins between 1988 and 2000 (Yang *et al.*, 2007; 2009). We  
356 observe that snow mass represents the major part of the TWS. The discharge peak is  
357 observed in June except for the Severnaya Dvina basin where the discharge is maximum  
358 in May.  
359

##### 360 ***4.2 Estimation of the correlation and time lag between SWE, TWS and river discharge***

361  
362 To determine which reservoir, snow or total water, has the most significant effect  
363 on river discharge, we computed the cross-correlation function between the time series of  
364 TWS and snow component and the time series of integrated discharge for each drainage  
365 basin when river discharges are available. We estimated correlation between the time  
366 series of snow volume and the time series of integrated discharge for each basin, using  
367 the cross-correlation function (equation 8) and the time lag between snow and discharge  
368 peaks corresponds to the month  $t_0$  that maximizes the cross-correlation function  
369 (equation 9).

370 The results obtained are presented in table 4 for the maximum of correlation and  
371 the time lag between the peak of land waters or snow and the peak of discharge. A good

372 agreement between snow storage derived from GRACE and discharge is observed for all  
373 the basins with correlation coefficients generally greater than 0.5 (table 4). For some  
374 basins, such as Lena, Mackenzie and Ob, the correlation is greater than 0.7. The  
375 correlation between TWS based on GRACE observations and discharge is lower  
376 whatever the basin you consider. For all the basins except Lena, correlations between  
377 snow mass and discharges and TWS and discharges are very close (the ratio between the  
378 correlation coefficients is greater than 0.75). In the case of the Lena basin, the correlation  
379 between snow storage from GRACE and river discharge is almost twice greater than the  
380 correlation between land water storage from GRACE and river discharge. These results  
381 are in accordance with the strong correlation observed between runoff and P-ET in the  
382 Lena watershed and the low correlation in the Mackenzie, Ob and Yenisey basins  
383 (Serreze *et al.*, 2003).

384 The time lag between snow mass, TWS and river discharge is an important  
385 variable for describing the snow-runoff relationship as the snow stored during winter is  
386 not a direct indicator of the river flow during summer. Different hydrological can affect  
387 snow: after melting, the snow can be evaporated, released as discharge, integrated to the  
388 interannual storage in ponds and wetlands (Bowling *et al.*, 2003).

389 The results obtained seem to be consistent with rivers morphology: long time lags  
390 (greater than 3 months) are obtained for large drainage basins such as Lena, Mackenzie,  
391 Ob and Yenisey, shorter time lags for smaller basins as Kolyma, Pechora and Severnaia  
392 Dvina. Some differences on the estimated time lags can be seen among the different  
393 datasets. They never exceed 1 month and can be caused by the monthly time sampling of  
394 the datasets. The results are in accordance with those obtained using SSM/I by Grippa *et al.*  
395 (2005) over the 1989-2001 period which found a strong correlation between snow  
396 depth in February and runoff in June, for the Ob basin, consistent with the time lag of (19  
397  $\pm$  7) pentads between peak of snow volume and maximum discharge for the Mackenzie  
398 basin over 1972-2000 (Dyer *et al.*, 2008), of 15 or 16 weeks for the Ob basin and of 16 to  
399 17 weeks for the Yenisey and Lena basins, between peaks of SWE derived from passive  
400 microwave observations and discharges over 1988-2000 (Yang *et al.*, 2007).

#### 401 **4.3 Interannual variability of GRACE-derived SWE**

402  
403 The GRACE-derived SWE interannual variability has been analyzed at basin  
404 scale. Maximum SWE has been estimated and compared to total annual discharge when  
405 the data are available. The results are presented on figure 8 for the Ob, Yenisey, Lena and  
406 Severnyy Dvina basins where data are available between 2003 and 2006. On the Western  
407 part of the Eurasian continent, *i.e.*, Severnyy Dvina, Pechora and Ob basins, a decline of  
408 both maximum SWE and total annual discharge. On the Eastern part of Eurasia, the  
409 increase of SWE during winter 2004 is followed by a decrease in 2005. If a good  
410 agreement with river discharge is observed for the Lena basin, the increase of the total  
411 discharge increases one year before the increase of SWE in the Yenisey basin. This  
412 difference of behaviour is probably caused by the effect of melt of permafrost (which  
413 covers 90% of the surface of the Yenisey basin) and the influence of the dams on the  
414 seasonality of the discharge is strongest in this basin than in other Eurasian basins  
415 (McClelland *et al.*, 2004). In the Mackenzie basin (not shown), the mean winter SWE and  
416 annual river discharge present a similar time evolution: a decrease between 2003 and

417 2004 followed by an increase in 2005. The **maximum** SWE is decreasing between 2005  
418 and 2006. In the Yukon basin (not shown), the **maximum** SWE remains constant between  
419 2003 and 2004, before increasing in 2005 and decreasing in 2006.

420

#### 421 **4.4 Basin-scale comparisons between GRACE-derived TWS and P-E**

422

423 The GRACE-derived TWS estimates can be compared to E-P through the  
424 instantaneous equation of the water mass balance applied to a watershed (see Hirschi *et*  
425 *al.*, 2006, for instance):

$$426 \quad \frac{\partial W}{\partial t} = P - ET - R \quad (10)$$

427 where  $\frac{\partial W}{\partial t}$ ,  $P$ ,  $ET$ ,  $R$  are water mass storage, precipitation rate, evapotranspiration rate  
428 and runoff respectively. Time integration of equation 10 between times  $t_1$  and  $t_2$  (the  
429 starting and the ending dates of the considered period, with  $\Delta t = t_2 - t_1$ , assumed to be ~30  
430 day, the average time span over which the GRACE geoids are provided) gives:

$$431 \quad \Delta W = \Delta P - \Delta ET - \Delta R \quad (11)$$

432 where  $\Delta W$ ,  $\Delta P$ ,  $\Delta ET$ ,  $\Delta R$  are the monthly changes of the parameters of equation 10.

433 As no **gridded** runoff data were available for the Arctic region for the study  
434 period, we directly compare TWS changes with P-E for the 11 largest Arctic basins and  
435 cannot determine if the water budget is closed (*i.e.*, Equation 10 fully verified). Time  
436 series of monthly changes of TWS and P-E are presented in figure 9. They generally  
437 present a similar evolution and range with respect to time, peaking during Northern  
438 Hemisphere autumn and reaching minima in May or June, in good accordance with  
439 climatologies (Serreze *et al.*, 2003), except for Nelson where P-E is lower than TWS  
440 change. The strong negative P-E anomaly of summer 2004, seen in most of the basins is  
441 well-observed in GRACE-derived TWS change.

442 Correlations between TWS change and P-E were computed and are reported in  
443 table 5. They allow us to determinate which fluxes influence the most the TWS change.  
444 Three types of Arctic basins can be distinguished : (1) a strong influence of P-E on TWS  
445 change (*i.e.*, correlation between TWS and P-E greater than 0.55) for the Ob, Yenisey  
446 and **Mackenzie** basins, (2) similar effect of P-E and runoff on TWS change (*i.e.*,  
447 correlation between TWS and P-E greater than 0.5) for the Yukon, Severnaia Dvina and  
448 Pechora basins, and (3) runoff dominates TWS change (*i.e.*, correlation between TWS  
449 and P-E lower than 0.3) for the Nelson, Kotya, Lena, Indigirka and Kolyma basins.  
450 Similar results were also found by Serreze *et al.* (2003) for the Ob, Yenisey, Lena and  
451 Mackenzie basins, comparing climatologies of P-ET and runoff.

#### 452 **4.6 Trends of SWE, TWS and river discharges**

453

454 Basin-scale trends of SWE, TWS, river discharges and PGR were estimated over  
455 2003-2006 using equation 5b. **As the hydrological signals are not stationary, these trends**  
456 **are valid over the 2003-2006 period.** The results are presented in table 6. PGR represents  
457 a possible source in our SWE and TWS trend estimates as GRACE, which measures  
458 vertically-integrated gravity, cannot distinguish between snow/water and other solid

459 Earth signals. We used the global ICE-5G deglaciation model of Peltier (2004), modified  
460 by Paulson *et al.* (2007), to compute PGR trends in each watershed (cf. table 6). PGR has  
461 a very important effect over Canada (18.8 and 25.6 km<sup>3</sup>.yr<sup>-1</sup> for Nelson and Mackenzie  
462 basins respectively) and also, but not such importantly and with the opposite sign, the  
463 large Siberian basins (-0.96, -0.77 and -0.59 km<sup>3</sup>.yr<sup>-1</sup> for Ob, Yenisey and Lena basins  
464 respectively. For the other basins, the effect of PGR is lower than 0.3 km<sup>3</sup>.yr<sup>-1</sup> in absolute  
465 value). In case of error-free PGR modelling, the PGR effect should be subtracted from  
466 SWE and TWS trends. Unfortunately, PGR effects remain not so well modelled since  
467 there are still large uncertainties on the of the Earth's interior (*i.e.*, constant of viscosity  
468 between upper and lower mantles for instance). So, we decided to present the results  
469 without PGR correction that suffers from important error itself (table 6). We notice that  
470 all the large Eurasian basins (except Indigirka where the inverse method gives unreliable  
471 results) considered in this study are losing snow mass, even in large amount as for Lena  
472 (-8.6 ± 0.9 km<sup>3</sup>.yr<sup>-1</sup>), Ob (-5.7 ± 1.4 km<sup>3</sup>.yr<sup>-1</sup>) and Severnaia Dvina (-4 ± 0.4 km<sup>3</sup>.yr<sup>-1</sup>). On  
473 the contrary, all the large North American basins are gaining snow mass (even Nelson  
474 where the inverse method gives unreliable results), except if we consider the effect of the  
475 PGR. Even if complex hydrological mechanisms occurred after the snow melt (Bowling  
476 *et al.*, 2003), increase (decrease) in snow mass could be explained by a decrease  
477 (respectively increase) of snow melt, and as a consequence by a decrease (respectively  
478 increase) of river discharges in North America (Eurasia respectively). This will be in  
479 accordance with the large increase of the Eurasian river discharges and the small decrease  
480 of the North American river discharges (Mc Clelland *et al.*, 2005). Nevertheless,  
481 comparisons with in-situ discharges data does not validate this assumption as Mackenzie  
482 discharge is increasing of (0.5 ± 0.5) km<sup>3</sup>.yr<sup>-1</sup> and, Ob and Severnaia Dvina discharges  
483 are respectively decreasing of (0.8 ± 0.5) km<sup>3</sup>.yr<sup>-1</sup> and (1.6 ± 0.2) km<sup>3</sup>.yr<sup>-1</sup> over the same  
484 period. For the Mackenzie basin, due to the effect of the PGR, the SWE trend should be  
485 negative and so, the assumption be verified. Trends of river discharges appear to be more  
486 correlated to trends in TWS. Moreover, Eurasia can be divided in two parts: the Western  
487 part with negative trends for SWE, TWS and river discharges and the Eastern part with  
488 negative trends for SWE but positive trends for TWS and river discharges. In the  
489 Northern Hemisphere, Siberia was one of the region the most affected by the recent  
490 warming (Jones and Moberg, 2003; NEESPI, 2004; Groisman *et al.*, 2006;  
491 <http://www.ncdc.noaa.gov/gcag/gcag.html>). This warming is responsible for both a melt  
492 of the snow and of the permafrost. On Western Eurasia where the permafrost only  
493 represents a small area and is discontinuous, TWS is mainly affected by decrease of  
494 SWE. On the contrary, on Eastern Eurasia where the permafrost is important, TWS could  
495 be recharged by a melt of the permafrost and TWS can increase whereas SWE decreases.

## 496 5. Conclusion

497 For the first time, GRACE-derived hydrological products are used to provide a  
498 description of several hydrological processes related to snow in the Arctic drainage  
499 system. Comparisons between GRACE-derived SWE, and snow depth and SWE  
500 climatologies show a relatively good agreement between the different products. Besides,  
501 comparisons between GRACE-derived SWE product and snowfall derived from GPCP  
502 rainfall over the same time period exhibit very similar spatial patterns (correlations of

503 0.75 have been reached for Yukon and [Mackenzie](#) basins) and interannual variations well  
504 correlated time variations at basin-scale. The direct comparison at basin scales between  
505 snow mass variations, land water variations and river discharge time evolution shows that  
506 the snow component has a more significant impact on river discharge at high latitudes  
507 than TWS, corroborating the results found earlier by Yang *et al.* (2003), and suggests a  
508 strong linkage between snow cover extent and streamflow. Time lags between snow mass  
509 maxima and discharge peaks, consistent with the size of drainage basins were also  
510 estimated. Interannual variability of SWE derived from GRACE is accordance with  
511 interannual variability of river discharges for most of the drainage basins. These results  
512 contribute to a better understanding of the relationship between snow and discharge and  
513 for hydrological models parameterization.

514 This study also provides a characterization of the respective influence of P-E and  
515 R on the TWS change at high latitudes. Time correlation between TWS and P-E allows to  
516 distinguish the drainage basins where TWS is dominated by P-ET, equally influenced by  
517 P-ET and runoff and the ones that remain mostly influenced by runoff.

518 The estimates of SWE and TWS trends over the 2003-2006 period showed that  
519 all the Eurasian basins loose snow mass whereas North American basins are gaining  
520 mass. Nevertheless, if we consider the effect of PGR, [Mackenzie](#) and Nelson basins also  
521 loose snow mass. Besides, Eurasia can be divided into two parts: the Western part where  
522 both SWE and TWS are decreasing and the Eastern part where SWE is decreasing  
523 whereas TWS is increasing. This different behavior could be related to melt of the  
524 permafrost mostly present in Eastern Eurasia.

525

526 **REFERENCES**

- 527 ADLER, R. F., HUFFMAN, G.J., CHANG, A., FERRARO, R., XIE, P., JANOWIAK,  
528 J., RUDOLF, B., SCHNEIDER, U., CURTIS, S., BOLVIN, D. GRUBER, A., J.  
529 SUSSKIND, J. and ARKIN, P., 2003, The Version 2 Global Precipitation Climatology  
530 Project (GPCP) Monthly Precipitation Analysis (1979-Present), *J. Hydrometeor.*, **4**, pp.  
531 1147-1167.
- 532  
533 ARCTIC CLIMATOLOGY PROJECT, 2000, Environmental working group Arctic  
534 meteorology and climate atlas, Edited by F. Fetterer and V. Radionov. Boulder, CO:  
535 National Snow and Ice Data Center. CD-ROM.
- 536  
537 ARCTIC RIMS, 2003, A Regional, Integrated Hydrological Monitoring, System for the  
538 Pan-Arctic Land Mass, available online at: <http://www.watsys.sr.unh.edu/arctic/RIMS/>.
- 539  
540 [BARNETT, T.P., ADAM, J.C., LETTENMEIER, D.P., 2005, Potential impacts of a](#)  
541 [warming climate on water availability in snow-dominated regions, \*Nature\*, \*\*438\*\*, 303-309,](#)  
542 [doi: 10.1038/nature04141.](#)
- 543  
544 BARRY R. G. and SERREZE, M.C., 2000, Atmospheric components of the Arctic  
545 Ocean freshwater balance and their interannual variability. In *The freshwater budget of*  
546 *the Arctic Ocean*. E. L. Lewis, et al. (Eds.), pp. 45-56, Kluwer Academic Publishers.
- 547  
548 BOWLING, L.C., KANE D.L., GIECK, R.E., HINZMAN L.D. and LETTENMEIER,  
549 D.P., 2003, The role of surface storage in a low-gradient Arctic watershed, *Water Resour.*  
550 *Res.*, **39(4)**, 1087, doi:10.1029/2002WR001466.
- 551  
552 BRASNETT B., 1999, A global analysis of snow depth for numerical weather prediction,  
553 *J. Appl. Meteorol.*, **38**, pp. 726-740.
- 554  
555 BROWN, R.D., BRASNETT, B. and ROBINSON, D., 2003, Gridded North American  
556 daily snow depth and snow water equivalent for GCM evaluation, *Atmosphere-Ocean*,  
557 **41**, pp. 1-14.
- 558  
559 [BULYGINA, O.N., RAZUVAEV, V.N., KORSHUNOVA, N.N., GROISMAN, P. YA.,](#)  
560 [Climate variations and changes in extreme climate events in Russia, 2007, \*Environ. Res.\*](#)  
561 [\*Let.\*, \*\*2\*\*, doi: 10.1088/1748-9326/2/4/045020.](#)
- 562  
563 CAO, Z., WANG, M., PROCTOR, B., STRONG G., STEWART R., RITCHIE, H. and  
564 BURFORD, J.E., 2002, On the physical processes associated with the water budget and  
565 discharge over the Mackenzie basin during the 1994/95 water year, *Atmosphere-Ocean*,  
566 **40(2)**, pp. 125-143.
- 567

568 CULLATHER, R.I., BROMWICH, D.H. and SERREZE, M.C., 2000, The atmospheric  
569 hydrologic cycle over the Arctic basin from reanalyses. Part I: Comparison with  
570 observations and previous studies, *J. Climate*, **13**, pp. 923-937.

571

572 DERY, S.J., SHEFFIELD, J. and WOOD, E.F., 2005, Connectivity between Eurasian and  
573 Canadian snow water equivalent and river discharge, *J. Geophys. Res.*, **110**, D23106, doi:  
574 10.1029/2005JD006173.

575

576 DOLL, P., KASPAR, F. and LEHNER, B., 2003. A global hydrological model for  
577 deriving water availability indicators: model tuning and validation, *J. Hydrol.*, **270**, pp.  
578 105-134.

579

580 [DYER, J., 2008, Snow depth and streamflow relationships in large North American  
581 watersheds, \*J. Geophys. Res.\*, \*\*113\*\*, D18113, doi:10.129/2008JD010031.](#)

582

583 FOSTER, D. J. and DAVY, R.D., 1988. Global snow depth climatology. *USAF  
584 ETAC/TN-88/006, Scott Air Force Base, Illinois*, 48 pp.

585

586 FRAPPART, F., RAMILLIEN, G., BIANCAMARIA, S., MOGNARD, N.M. and  
587 CAZENAVE, A., 2006, Evolution of high-latitude snow mass derived from the GRACE  
588 gravimetry mission (2002-2004), *Geophys. Res. Lett.*, **33**, L02501,  
589 doi:10.1029/2005GL024778.

590

591 GRIPPA, M., MOGNARD, N.M. and LE TOAN, T., 2005, Comparison between the  
592 interannual variability of snow parameters derived from SSM/I and the Ob river  
593 discharge, *Remote Sens. Env.*, **98**, pp. 35-44.

594

595 GROISMAN, P.Y., KNIGHT, R.W., RAZURAEV, V.N., BULYGINA, O.N. and  
596 KARL, T.R., 2006, State of the Ground: Climatology and Changes During the Past 69  
597 Years over Northern Eurasia for a Rarely Used Measure of Snow Cover and Frozen  
598 Land, *J. Climate*, **19** (19), pp. 4933–4955.

599

600 HIRSCHI, M., SENEVIRATNE, S.I. and SCHAR, C., 2006, Seasonal variations in  
601 terrestrial water storage for major midlatitude river basins, *J. Meteorol.*, **7**(1), pp. 39–60,  
602 doi:10.1175/JHM480.1.

603

604 JONES, P.D. and MOBERG, A., 2003, Hemispheric and large-scale surface air  
605 temperature variations: an extensive revision and an update to 2001, *J. Climate*, **16**, pp.  
606 206–223.

607

608 [KITAEV, L., FØRLAND, E., RAZUVAEV, V., TVEITO, O.E., KRUEGER, O., 2005,  
609 Distribution of snow cover over Northern Eurasia, \*Nordic Hydrol.\*, \*\*36\*\*\(4-5\), 311-319.](#)

610

611 LAMMERS, R.B., SHIKLOMANOV, A.I., VOROSMARTY, C.J., FEKETE, B.M. and  
612 PETERSON, B.J., 2001, Assessment of contemporary Arctic river runoff based on  
613 observational discharge records, *J. Geophys. Res.*, **106** (D4), pp. 3321-3334.



614  
615 KALNAY, E., KANAMITSU, M., KISTLER, R., COLLINS, W., DEAVEN, D.,  
616 GANDIN, L., IREDELL, M., SAHA, S., WHITE, G., WOOLEN, J., ZHU, Y.,  
617 CHELLIAH, M., EBISUZAKI, W., HIGGENS, W., JANOWIAK, J., MO, K.C.,  
618 ROPELEWSKI, C., WANG, J., LEETMA, A., REYNOLDS, R., JENNE, R. and  
619 JOSEPH, D., 1996, The NCEP/NCAR 40-year reanalysis project, *Bull. Amer. Meteorol.*  
620 *Soc.*, **77**, pp. 437-471.

621  
622 MC CLELLAND, J.W., HOLMES, R.M., PETERSON, B.J. and STIEGLITZ, M., 2004,  
623 Increasing river discharge in the Eurasian Arctic: Consideration of dams, permafrost  
624 thaw, and fires as potential agent of change, *J. Geophys. Res.*, **109**, D18102, doi:  
625 10.1029/2004JD005483.

626  
627 MITROVICA, J.X., WAHR, J., MATSUYAMA, I. and PAULSON, A., 2005. The  
628 rotational stability of an Ice Age Earth, *Geophys. J. Int.*, **161**, pp. 491-506.

629  
630 MOLNIA, B.F., 2007, Late nineteenth to early twenty-first century behavior of Alaskan  
631 glaciers as indicators of changing regional climate, *Global Planet. Change*, **56(1-2)**, pp.  
632 23-56.

633  
634 NIU, G.-Y., SEO, K-W. Yang, Z-L., WILSON, C., SU, H., CHEN, J. & RODELL, M.  
635 2007. Retrieving snow mass from GRACE terrestrial water storage change with a land  
636 surface model, *Geophys. Res. Lett.*, **34**, L15704, doi:10.1029/2007GL030413.

637  
638 NORTHERN EURASIA EARTH SCIENCE PARTNERSHIP INITIATIVE (NEESPI),  
639 2004. In *Science Plan*, Groisman, P.Y.a., Bartalev, S.A. (Eds.), Prepared by the NEESPI  
640 Science Plan Development Team. Available online at: <http://neespi.org>.

641  
642 OKI, T. and SUD, Y.C., 1998. Design of Total Runoff Integrating Pathways (TRIP) - A  
643 global river channel network. *Earth Interactions*, **2 (1)**, pp. 1-37.

644  
645 OSTERKAMP, T.E., 2005. The recent warming of permafrost in Alaska,  
646 *Global Planet. Change*, **49(3-4)**, pp. 187-202.

647  
648 PAULSON, A., ZHONG, S. and WAHR, J., 2007, Inference of mantle viscosity from  
649 GRACE and relative sea level data, *Geophys. J. Int.*, **171 (2)**, pp. 497-508.

650  
651 PELTIER, W.R., 2004, Global Glacial Isostasy and the Surface of the Ice-Age Earth: The  
652 ICE-5G(VM2) model and GRACE, *Ann. Rev. Earth Planet. Sci.*, **32**, pp. 111-149

653  
654 PETERSON, B.J., HOLMES, R.M., MC CLELLAND, J.W., VOROSMARTY, C.J.,  
655 LAMMERS, R.B., SHIKLOMANOV, A.I., SHIKLOMANOV, I.A., RAHMSTORF, S.,  
656 2002, Increasing river discharge to the Arctic Ocean, *Science*, **298**, pp. 2171-2173.

657

- 658 PROSHUTINSKY, A., POLYAKOV, I., JOHNSON, M., 1999, Climate states and  
659 variability of Arctic ice and water dynamics during 1946-1997, *Polar Res.*, **18(2)**, 135-  
660 142.
- 661  
662 RAMILLIEN, G., FRAPPART, F., CAZENAVE, A. and GUENTNER, A., 2005, Time  
663 variations of the land water storage from an inversion of 2 years of GRACE geoids, *Earth*  
664 *Planet. Sci. Lett.*, **235**, pp. 283-301.
- 665  
666 RAMILLIEN, G., FRAPPART, F., CAZENAVE, A., GUENTNER, A. and LAVAL, K.  
667 2006, Time variations of the regional evapotranspiration rate from Gravity Recovery and  
668 Climate Experiment (GRACE) satellite gravimetry, *Water Resour. Res.*, **42(10)**, W10403,  
10.1029/2005WR004331.
- 669  
670 RANGO, A., 1997, Response of areal snow cover to climate change in a snowmelt-runoff  
671 model, *Annals of Glaciology*, **25**, pp. 232-236.
- 672  
673 RODELL, M. and FAMIGLIETTI, J.S., 1999, Detectability of variations in continental  
674 water storage from satellite observations of the time dependent gravity field, *Water*  
675 *Resources Res.*, **35 (9)**, pp. 2705-2723.
- 676  
677 SERREZE, M.C., BROMWITCH, D.H., CLARK, M.P., ETRINGER, A.J., ZHANG, T.  
678 and LAMMERS, R.B. (2003). Large-scale hydro-climatology of the terrestrial Arctic  
679 drainage system, *J. Geophys. Res.*, **108**, 8160, doi: 10.1029/2001JD000919.
- 680  
681 STOCKER, T.F., and RAIBLE, C.C., 2005, Water cycle shifts gear, *Nature*, **434**, pp.  
682 830-833.
- 683  
684 SWENSON, S., WAHR, J. and MILLY, P.C.D., 2003, Estimated accuracies of regional  
685 water storage variations inferred from the Gravity Recovery and Climate Experiment  
686 (GRACE). *Water Resour. Res.*, **39 (8)**, 1223, doi:10.1029/2002WR001808.
- 687  
688 SYED, T. H., FAMIGLIETTI, J.S., ZLOTNICKI, V., and RODELL, M., 2007,  
689 Contemporary estimates of Pan-Arctic freshwater discharge from GRACE and reanalysis,  
690 *Geophys. Res. Lett.*, **34**, L19404, doi:10.1029/2007GL031254.
- 691  
692 TAPLEY, B.D., BETTADPUR, S., WATKINS, M., and REIGBER, C., 2004 a, The  
693 Gravity Recovery and Climate Experiment : Mission overview and Early results,  
694 *Geophys. Res. Lett.*, **31**, L09607, doi:10.1029/2004GL019920.
- 695  
696 TAPLEY, B.D., BETTADPUR, S., RIES, J., THOMPSON, P.F. and WATKINS, M.,  
697 2004 b, GRACE measurements of mass variability in the Earth system, *Science*, **305**, pp.  
698 503-505.
- 699  
700 WAHR, J., SWENSON, S., and VELICOGNA, I., 2006, The Accuracy of GRACE Mass  
701 Estimates, *Geophys. Res. Lett.*, **33**, L06401, doi:10.1029/2005GL025305.

- 695 YANG, D., ROBINSON, D., ZAO, Y., ESTILOW, T., YE, B., 2003, Streamflow  
696 response to seasonal snow cover extent changes in large Siberian watersheds, *J. Geophys.*  
697 *Res.*, **108(D18)**, 4578, doi: 10.1029/2002JD003419.
- 698 YANG, D., ZHAO, Y., ARMSTRONG, R., ROBINSON, D., BRODZIK, M.J., 2007,  
699 Streamflow response to seasonal snow cover mass changes over large Siberian  
700 watersheds, *J. Geophys. Res.*, **112**, F02S22, doi: 10.129/2006JF000518.
- 701 YANG, D., ZHAO, Y., ARMSTRONG, R., ROBINSON, D., 2009, Yukon river  
702 streamflow response to seasonal snow cover changes, *Hydrol. Process.*, **23**, 109-121, doi:  
703 10.1002/hyp.7216.

704 **Acknowledgements**

705 The first and third authors were supported by NASA GRACE Science Team grant  
706 NG04GE99G and NASA REASoN grant JPL 1259524.  
707

708 **Table caption**

709

710 [Table 1](#): Surface of the largest Arctic drainage basins (ArcticRIMS, 2003), mean annual  
711 discharges (CAFF, 2001), and availability of the discharges data at the closest station to  
712 the mouth. (source: ArcticRIMS, 2003).

713

714 [Table 2](#): Spatial and temporal resolutions, and period acquisition of the GRACE-derived  
715 SWE and of the datasets used for comparisons.

716

717 [Table 3](#): Correlation and time lag between GRACE-derived SWE and snowfall derived  
718 from GPCP by river basin.

719

720 [Table 4](#): Correlation and time lag between fresh water volume and snow stored for the  
721 different remote sensing datasets by river basin.

722

723 [Table 5](#): Correlation between TWS and P-E by river basin.

724

725 [Table 6](#): Trends of snow volume, TWS volume and water volume to the Arctic Ocean  
726 (when data are available) estimated between 2003 and 2006 by river basin.

727

728

729 **Figure caption**

730 Figure 1: Location of the main Arctic drainage basins and their annual discharge ( $\text{km}^3\text{yr}^{-1}$ ). Source: Major River Systems in the Arctic. UNEP/GRID-Arendal Maps and Graphics  
731 Library. 2002. Available at:  
732 [http://maps.grida.no/go/graphic/major\\_river\\_systems\\_in\\_the\\_arctic](http://maps.grida.no/go/graphic/major_river_systems_in_the_arctic).  
733

734 Figure 2: Maps of SWE amplitude of annual cycle derived from GRACE over the 2003-  
735 2006 period (a), mean annual snow depth from the ETAC monthly snow climatology (b),  
736 mean annual snow depth from the EWG monthly snow climatology (c) and total annual  
737 snowfall derived from GPCP rainfall over the 2003-2006 period (d).

738 Figure 3: Month where is maximum the GRACE-derived SWE (a), the snow depth from  
739 USAF/ETAC climatology (b), and the snow depth from EWG climatology (c).Figure 4:  
740 Time series of SWE (mm) derived from GRACE (continuous black) and snowfall (mm)  
741 derived from GPCP (dashed black) for the four largest Arctic drainage basins: Ob,  
742 Yenisey, Lena, Mackenzie.

743 Figure 4: Maps of SWE annual cycle derived from GRACE during the 2003-2006 period  
744 (a), mean annual SWE snow climatology (b) over North America. Corr=0.58.

745 Figure 5: Time series of SWE (mm) derived from GRACE (continuous black) and  
746 snowfall (mm) derived from GPCP (dashed black) for the four largest Arctic drainage  
747 basins: Ob (a), Yenisey (b), Lena (c), Mackenzie (d).

748 Figure 6: Time series of SWE (mm) derived from GRACE (black) and WGHM (dotted  
749 black) for the Nelson (a) and Indigirka (b) basins.

750 Figure 7: Time series of SWE (mm) derived from GRACE (red), of TWS derived from  
751 GRACE (black) and of river discharges (blue) for the six Arctic drainage basins: Ob (a),  
752 Yenisey (b), Lena (c), Mackenzie (d), Kolyma (e) and Severnaya Dvina (f).

753 Figure 8: Time series of annual maximum of SWE (mm) derived from GRACE (black)  
754 and total annual discharge ( $10^3 \text{ m}^3\cdot\text{s}^{-1}$ , dotted black) for the Ob (a), Yenisey (b), Lena (c)  
755 and Severnaia Dvina (d) basins.

756 Figure 9: Time series of monthly TWS change (mm/month) derived from GRACE  
757 (black) and P-E (dotted black) for the Ob (a), Yenisey (b), Lena (c), Mackenzie (d),  
758 Nelson (e) and Yukon (f) basins.