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Modeling the water resources of the Black and Mediterranean Sea river basins and their impact on regional mass changes

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Abstract

For the first time, a dedicated release of the hydrology and water use model WaterGAP3, has been developed to spatially explicit calculate hydrological fluxes within river basins draining into the Mediterranean and Black Sea. The main differences between the new regional version of the global WaterGAP3 model and the previously applied global version WaterGAP2 can be found in the spatial resolution, snow modeling, and water use modeling. Comparison with observations shows that WaterGAP3 features a more realistic representation of modeled river runoff and inflow into both seas. WaterGAP3 generates more inflow to both seas than WaterGAP2. In the WaterGAP3 simulation, contributions to the total runoff into the Black Sea from individual discharge regions show in general a good agreement to climatology derived runoff, but lesser importance of Georgian rivers for the basin's water. After the successful model validation WaterGAP3 has been applied to correct estimates of seawater mass derived from the GRACE gravity mission and to account for freshwater inflow into both basins. The performance of the WaterGAP3 regional solution has been evaluated by comparing the seawater mass derived from GRACE corrected for the leakage of continental hydrology, to an independent estimate derived from steric-corrected satellite altimetry with steric correction from regional oceanographic models. The agreement is higher in the Mediterranean Sea than in the Black Sea. Results using WaterGAP3 and WaterGAP2 are not significantly different. However the agreement with the altimetry-derived results is higher using WaterGAP2, due to the smaller annual amplitude of the continental hydrology leakage from WaterGAP3. We conclude that the regional model WaterGAP3 is capable of realistically quantifying water mass variation in the region, further developments have been identified.

Keywords: GRACE, mass transport, oceanography, altimetry, WaterGAP

1 Introduction

In terrestrial hydrology, the separation of water in the numerous storage compartments is a highly complex problem. The heterogeneous nature of soil and topography results in water storage variations which can vary strongly in the spatial domain. At the same time, precipitation, evapotranspiration, water routing and anthropogenic water use all play a strong role in the (re-)destribution of water. Often depending on integrative measurements, such as observed river runoff, the use of models in hydrology is indispensable in understanding the (re-)destribution of water. Their utilization is twofold; on the one hand, they provide estimates of water storage where observations alone do not suffice, while on the other hand they provide an opportunity to study the physics of the system by means of simulations. The validation/calibration of the hydrological models depends on independent observations. River gauges are traditionally used to calibrate the models, but the data is sparse or unavailable. Since the launch of the Gravity Recovery and Climate Experiment (GRACE), large scale measurements of total water storage have been available (Schmidt et al. 2006). When used in conjunction with in situ observations, GRACE data have the potential to significantly improve our understanding of hydrological processes and the global water cycle and provide opportunities for model calibration (Werth 2010). A comprehensive overview of hydrological and land surface models dealing with GRACE derived mass variations is given by Güntner (2008).

In this study, a regional version of the global hydrology and water use model WaterGAP3 (Aus der Beek et al. 2011) has been developed, which includes all river basins draining into the Mediterranean and Black Sea. In addition to the increased spatial resolution (5 arc minutes), special attention was given to improved process algorithms (e.g. snow hydrology and flow velocity) and more detailed anthropogenic water use modules, which is a significant factor in the heavily populated drainage basins (Aus der Beek et al. 2010). In order to evaluate the performance of WaterGAP3 and the goodness of its results, modeled river runoff has been compared to observed river runoff from gauging stations available in the study region. Furthermore, to grade the new regional WaterGAP3 model version, its results have also been compared to those from WaterGAP2 and a land surface model. After the successful verification of the WaterGAP3 model results, monthly maps of total water storage (TWS) have been produced. We then use this TWS data to correct the water mass variation estimated by GRACE (Chambers 2006, Fenoglio-Marc et al., 2006, 2007). On the other hand, also steric-corrected altimetry gives an estimation of seawater mass. The consistency between the two estimates is affected by the continental hydrological leakage and steric corrections (Fenoglio-Marc et al. this issue), therefore the modeled TWS can be evaluated through the comparison. Here, we compare steric-corrected altimetry with GRACE corrected by using the regional WaterGAP3 and the global WaterGAP2 models.

2 Material and Methods

2.1 Regional modeling of hydrology using WaterGAP3

The hydrology and water use model applied in this study is WaterGAP3 (Water – Global Assessment and Prognosis). It is a further developed version of the well known global hydrology and water use model WaterGAP2 (Alcamo et al. 2003, Döll et al. 2003). WaterGAP3 operates on the global scale but for each continent separately, whereas a special landmask has been developed for this study, as the study region includes parts of three continents (Europe, Africa, and Asia). Thus, the landmask includes all river basins draining into the Black and Mediterranean Sea. In addition to the increase in spatial resolution from 0.5° (~50 km x 50 km) to 5’ (~7 km x 7 km) (Verzano 2009) also some hydrological process descriptions have been improved in WaterGAP3, such as snow (Verzano & Menzel 2009) and permafrost modeling (Aus der Beek & Teichert 2008), flow velocity (Verzano et al. 2005), and the water use modules (Aus der Beek et al. 2010; Flörke et al. 2011).

WaterGAP3 calculates daily water fluxes and anthropogenic water abstractions for all river basins draining into the Mediterranean and Black Sea (see Figure 1) on a 5 by 5 arc minutes grid (longitude and latitude). For each grid cell it takes into account spatially distributed physiographic information about elevation, slope, hydrogeology, land cover and soil properties, as well as location and extent of lakes, wetlands, and reservoirs. The upstream/downstream relationship among the grid cells is defined by a global drainage direction map (DDM) which indicates the drainage direction of surface water (Lehner et al. 2008). Five water use models within the WaterGAP framework (Flörke and Alcamo 2005, Alcamo et al. 2003, Döll and Siebert 2002) take into account water consumption by households, manufactures, energy production industries, livestock, and irrigation, as the consideration of water uses is crucial for a realistic representation of the water balance of most river basins. The aim of WaterGAP3 is to simulate the characteristic behaviour of the terrestrial water cycle in order to estimate total water storage (TWS) for each grid cell and river basin. Herein, TWS is defined as the sum of all potential water storages: rivers, groundwater, soil water, canopy water, snow, lakes, and wetlands. River runoff is calibrated and validated with a single tuning parameter (Döll et al. 2003) against 63 stations of observed river flow (GRDC 2004). Climate forcing data used in this study have been compiled from station data and regionalized by the Climate Research Unit (CRU) of the University of East Anglia, Norwich, U.K. (versions TS 1.2 and TS 2.1, Mitchell and Jones, 2005). CRU data covers the time period from 1901 to 2002 in 10' and 0.5° resolution and monthly time steps, providing nine climatic parameters, e.g. precipitation, air temperature, cloud cover, etc.. For the time period 2003 to 2009 climate forcing data has been provided by the ECMWF (European Centre for Medium Weather Forecast) in 0.35° (until January 2006) and 0.22° resolution. Precipitation data for this time period originates from the GPCC (Global Precipitation and Climate Center) and features a spatial resolution of 0.5° (until December 2007) and 1° (Rudolf & Schneider 2005). All climate input data has been converted to the 5' WaterGAP3 grid. Each 5' cell is linked to a, e.g., 0.5 cell based on geographical location of the center of the 5' cell. Thus, the climate data has only been remapped to the 5' grid, and not re-interpolated.

Hydrological modeling has been conducted for the time period 1901 to 2009. This ensures an optimal utilization of station data for the calibration of WaterGAP3, as observed runoff data generally is scarce after 2002. Only river basins draining into the Mediterranean and Black Sea have been selected for this regional version of WaterGAP3 (see Figure 1). River basins, which do not drain into the Black or Mediterranean Sea, but which are located within other river basins that actually do drain into the Black or Mediterranean Sea, such as in Turkey, the Balkans, or in the Nile basin, are considered as inland sinks. Thus, they are excluded from the model set-up.

To test the performance of the regional WaterGAP3 model, its results are being compared to the results from the models WaterGAP2 and GLDAS (Global Land Data Assimilation System) (Rodell et al. 2004). GLDAS drives four different land surface models (LSMs): CLM, MOSAIC, NOAH, VIC which simulate the transfer of mass, energy, and momentum between the soil and vegetation surfaces and the atmosphere. In this study, routed river runoff data from all four LSMs has been provided for the Danube and Dnieper rivers (Zaitchik 2010). To simplify the comparison an unweighted mean of all four GLDAS LSMs has been applied. Furthermore, the WaterGAP3 results are also being compared to values from literature (Ludwig et al. 2009; Jaoshvili 2002) and climatological estimates (Grayek et al 2010).

To test the performance of the regional WaterGAP3 model in computing the continental hydrology leakage on GRACE, the global models WaterGAP2, Land Dynamics (LaD) Fraser (Mill and Shmakin, 2002) and Community Land Model of the Global Data Assimilation System (GLDAS-CLM) (Rodell et al., 2004) have been used.

2.2 GRACE gravimetry and Jason-1 multi-satellite altimetry

In order to assess the influence of the hydrological models as corrections to the basin averaged ocean mass estimated from GRACE, we have constructed time series from GRACE gravimetry on the Mediterranean and on the Black Sea basins. The hydrological leakage has been estimated by converting the total water storage from the hydrological models to the spectral domain and computing monthly filtered basin averages in the same way as for GRACE. The general procedure is outlined below, more details on the processing can be found in Fenoglio-Marc et al. (this issue).

We have used the GRACE data (GFZ level-2 product, release 4) after restoring the non-tidal oceanic and atmospheric background models (GAD) over ocean. The seasonal degree 1 terms of the geopotential have been included (Rietbroek et al., 2009, 2011), the low degree coefficients replaced with a more accurate estimate from satellite laser ranging (SLR) and the high coefficients truncated to degree 100. Basin means over each ocean basin (defined as 1 on sea and 0 on land) have been estimated applying a dedicated anisotropic post-processing filter DDK3 (Kusche et al. 2009, Swenson and Wahr 2002). The same procedure has been applied to observed and model data, e.g. to GRACE and to the hydrology model. The filtered basin average of the hydrological signal is then removed from the GRACE filtered basin average to obtain the hydrological-corrected seawater mass change from GRACE.

Steric-corrected sea level derived from multi-satellite (Jason-1 and Envisat) altimetry data have been used to obtain an independent estimate of the ocean mass signal. To correct for the steric component, maps of the steric component of sea level have been derived from the oceanographic models MFSTEP (Dobricic & Pinardi 2008) in the Mediterranean Sea and NEMO (Madec 2008) in the Black Sea (Grayek et al. 2010). Filtered basin averages have been constructed with the same method applied to GRACE and a re-scaling factor derived from the filtered and unfiltered mass estimated from steric-corrected altimetry (1.4 for the Mediterranean Sea and 1.7 for the Black Sea, see Fenoglio-Marc et al., this issue). Finally, the re-scaling factors have been applied to the hydrology-corrected seawater mass change estimated from GRACE. The difference between the steric-corrected altimetry and the hydrology-corrected GRACE seawater mass estimates allows an evaluation of the steric- and continental hydrology leakage corrections.

3. Results

3.1 Hydrological model results and evaluation

3.1.1 Total water storage

Mean TWS, as calculated with WaterGAP3 for the period 2002 to 2009, is displayed in Figure 1. The black borders show the extent of the 48 largest basins draining into the Mediterranean and Black Sea. All these basins feature an area greater than 1500 km², as this size has been identified to be the minimum for producing reliable results with WaterGAP3. The highest TWS can be found in the Alps and Balkan Mountains, as well as in the northern part of the Dnieper and Don basins, and at the eastern coast of the Black Sea. Mountainous regions generally provide larger TWS as higher precipitation amounts cause higher filling levels of storages. In addition, lower air temperatures in high elevations entail long duration snow storages. Furthermore, large rivers, such as the Nile, Danube, and Dnieper, also feature high TWS in their river beds and riparian zones. Small TWS generally occur in arid to semi-arid regions, such as Northern Africa, but also in areas with very small hydrological storage compartments, as for example in large coastal parts of Greece. TWS can strongly be affected by anthropogenic water abstraction and consumption, which can also alter the hydrological runoff regimes of rivers. Exemplary, Figure 2 shows mean annual water withdrawals for irrigation purposes. Hot spots with high intensity and large extents of irrigation systems can be found in the Nile Delta, the Italian Po basin, and the Spanish Ebro basin. The Ebro basin features a basin wide average TWS of 83.5 mm, whereas on average 54.7 mm have already been withdrawn during the model application for irrigation requirements. Taking into account a Spain-explicit irrigation project efficiency of 53% (aus der Beek et al. 2010), 29 mm have been consumed by plants and 25.7 mm have remained as return flow in the Ebro basin. This means that without considering irrigation water uses, TWS would have been calculated as 112.5 mm, which again points out the importance of considering water uses when modeling hydrological water balances.

3.1.2 Modeled river runoff to the Mediterranean and Black Sea

Figure 3 shows monthly time series of cumulated modeled river runoff into the Black Sea. WaterGAP3 generally produces higher peak flows as WaterGAP2, whereas the timing and phase are similar. Peak flows are dominated by the pluvio-nival regime of the Danube River, which features snowmelt- and intense rainfall induced high flows between March and May. The Danube contributes 51% of the total inflow to the Black Sea, followed by Don (13.9%), Dnieper (11.4%), Kuban (6.8%), Rioni (3.7%), Coruh (3.1%), Dniester (2.8%), Kizil Irmak (2.1%), Sakarya (1.9%), Yesil Irmak (1.8%), Filyos (1.0%) and Bug (0.8%). A comparison with Dai et al. (2009) and a literature review conducted by Ludwig et al. (2009) yields a range of inflow into the Black Sea of 348 to 473 km^3 per year. WaterGAP3 calculated an average inflow for 2002 to 2009 of 406 km^3 per year, which is close to the 403 km^3 per year estimated by Ludwig et al. (2009) for the period 1991 to 2000 (WaterGAP3: 337 km^3 per year). WaterGAP2 simulates an average cumulated inflow into the Black Sea of 326 km^3 per year, which is below the literature review based range mentioned above. This is expected to be caused by the generally small base flow produced in late summer. Differences in the results of the models can be explained by the differences in model structures and drivers mentioned in chapter 2.1. Especially, the spatial sub-grid resolution of 400 m x 400 m of the WaterGAP3 snow module reacts more dynamically and sensitive to air temperature changes, compared to WaterGAP2 with a spatial resolution of 50 km x 50 km (Verzano & Menzel 2009). The higher spatial resolution can cause higher snowmelt induced peak flows, as the sub grid cells provide individual snow storage volumes based on elevation, albedo and climatic drivers. In general, WaterGAP2 produces 19% less inflow to the Black Sea than WaterGAP3. Potential sources for this discrepancy might be found in the application of different precipitation inputs, as well as in the station correction factor of WaterGAP2, which can lead to negative water balances, as it manipulates the water balance in order to fit modeled to observed river runoff. Furthermore, WaterGAP2 applies constant water withdrawals for 2002 to 2009, which can lead to the overestimation of water consumption, especially in dry seasons when precipitation amounts are low. The allocation of irrigated areas in WaterGAP2 relies on the concept of irrigating areas which are potentially equipped for irrigation (Döll & Siebert 2002), whereas in WaterGAP3 irrigated areas have been adjusted to national statistics and reported areas (Aus der Beek et al. 2010). Thus, irrigation water withdrawals are likely to be overestimated by WaterGAP2. This might also explain the lower base flow in late summer modeled with WaterGAP2, as most water withdrawals for irrigation purposes occur in hot and dry summer months.

Figure 4 depicts monthly time series of cumulated modeled river runoff to the Mediterranean Sea. Similar to the inflow to the Black Sea (see Figure 3), peaks modeled with WaterGAP3 are higher than from WaterGAP2, whereas the latter generates lower base flows. The river basins draining into the Mediterranean Sea are generally smaller than those of the Black Sea (see Figure 1). Here, the advantages of the fine spatial resolution of WaterGAP3 take effect. Most river basins are smaller than 15000 km², which is thought to be the minimum basin size for WaterGAP2 to yield realistic results; as its cell areas are approximately 2500 km² per grid cell. Thus, WaterGAP3, with a cell size of about 50 km², offers the ideal setting for a comprehensive model study for the Mediterranean River basins. WaterGAP3 calculates a mean annual cumulated inflow to the Mediterranean Sea of 518 km³ per year for the period 2002 to 2009, which is in the middle of the literature review based range of Ludwig et al. (2009) of 305 to 737 km³ per year. WaterGAP2 simulates about 337 km³ per year, whereas similar reasons as for the difference between both models in the Black Sea basins apply. The hydrological and water use system of the Nile River is very difficult to represent in a large scale hydrological model, as no data about river runoff, water withdrawals and dam management is available for the downstream area and Delta of the Nile. In detail, the reservoirs Aswan High Dam and Lake Nasser heavily influence the river runoff, and high intense double to triple cropping irrigation projects in the Nile Delta cause massive water withdrawals. In addition, the Egyptian part of the Nile River provides water to more than 121 million people, as 95% of all Egyptians live within 20 km distance of the Nile River. These water withdrawals have been included in this study. However, as no reported data about domestic water withdrawals in Egypt is available, it is difficult to verify these WaterGAP3 model results.

Modeled natural flows without anthropogenic interaction exceed 300 km³ per year, whereas literature values for present Nile runoff into the Mediterranean Sea range between 6 (Abu El Ella 1993) and 15 (Nixon 2003) km³ per year. These large discrepancies can currently not realistically be modeled with WaterGAP3. However, the TWS from Figure 1 are still realistic for the Nile, as a comparison of modeled with observed runoff in the Blue and White Nile sub-basins feature a very high level of agreement (R^2 = 0.89; Model efficiency = 0.68 at GRDC station “Roseires”). Therefore, in many studies as well as in Figure 2 the Nile is excluded from the cumulated inflow to the Mediterranean Sea. Without the Nile, WaterGAP3 simulates an average cumulated inflow for 2002 to 2009 of 318 km³ per year, which is close to the 312 km³ per year Ludwig et al. (2009) have reported for the period 1991 to 2000. Generally, WaterGAP3 generates 35% more inflow to the Mediterranean Sea than WaterGAP2. This can again be explained by similar reasons as for the Black Sea because the alpine river catchments with snowmelt and rainfall induced floods dominate the total inflow to the Mediterranean Sea. The Po River accounts for 24.2% of the cumulated inflow, followed by the Rhone (23.6%), Ebro (6.1%), Drin-Buna (4.1%), Orontes (3.6%), Maritsa (3.6%), Adige (3.5%), Moutouya (3.5%), Simav (3.2%), and other smaller rivers which individually contribute less than 3%.

3.1.3 Comparison of WaterGAP3 with other hydrological models and data

The results of the river runoff provided by WaterGAP3 for the Black Sea have been compared in terms of the total amount to those from semi-climatological estimates (Grayek et al 2010) and in terms shares from individual drainage regions to values from literature (Jaoshvili 2002). The comparison of the contributions from surrounding countries and correspondingly assigned WaterGAP3 river runoff are shown in Figure 5 and 6, respectively. As expected, the annual and inter-annual variation of total river runoff derived from the semi-climatological estimate (Figure 5) is much smaller than the variation simulated by WaterGAP3. Furthermore, the comparison of annual mean values from both datasets reveals that the simulated data gives in general higher values than the estimated data. The difference here reaches from 1500 to 4000 m³/s. In Figure 6, the contributions of the individual drainage regions from surrounding countries are shown. With one exception, which is the runoff of Georgia's rivers, the simulated percentage shares from drainage areas vary roughly around the climatological value from Jaoschvili (2002). The share of Georgian rivers in WaterGAP3 is significantly lower than in the climatology and, correspondingly, the temporal mean values of the other drainage regions are slightly higher than the climatological one. This can be explained by anthropological effects, such as reservoir management and water withdrawals, e.g. in the Kuban River basin (Ludwig et al .2009).

Figure 7 shows a comparison of modeled river runoff from three different hydrological models with observed river runoff at the GRDC Station “Ceatal Izmail” (ID 6742900) of the

Danube River. This station has been selected from the 63 stations of this study, as routed river runoff from GLDAS was only available for this station and for a station at the Dnieper River. However, river runoff at the Dnieper station is heavily regulated, which makes it difficult to judge the goodness of three models in this study. The performance of the models is evaluated by calculating the coefficient of determination R^2 and the Nash-Sutcliffe efficiency NSE (Nash & Sutcliffe 1970). WaterGAP3 generally yields the best fit of all three models (R^2 = 0.67, NSE = 0.50). Especially, the observed volumes of peak flows are very well represented by WaterGAP3, whereas the timing and phase also show a good fit to observed values. The second model, WaterGAP2, features a somewhat lower goodness of fit (R^2 = 0.47, NSE = -1.24), which again is caused by the underestimation of base flows. In particular, during the flood event of winter 1993/1994, WaterGAP2 simulates a strong decline in runoff, whereas observed runoff increases to the highest flood event of the year. This may be explained by rain-on-snow conditions, which can easily set free large amounts of water, due to rapid melting snow packs and frozen soil inducing Hortonian overland flow. These events are very difficult to be represented in a model, especially with a spatial resolution of 0.5°. The last model, GLDAS, generates runoff which is within a reasonable range (R^2 = 0.06, NSE = -2.49) in terms of runoff volume, whereas the inaccurate timing and phase of modeled runoff point at model structure deficiencies. However, it needs to be mentioned that both WaterGAP models have been calibrated with one simple parameter (Döll et al. 2003) to observed runoff, whereas all four GLDAS LSMs have not been calibrated.

A comparison of the goodness of WaterGAP2 and WaterGAP3 with observed data from all river discharge gauging stations available in the Mediterranean and Black Sea basins is shown in Figure 8. For each year the correlation between modeled and observed river runoff (in m^3) has been calculated separately for each station on a monthly basis. Then, an annual unweighted arithmetic mean from all available stations has been calculated. The comparison shows generally higher coefficients of determination for WaterGAP3 in the range of 0.50 to 0.54, whereas WaterGAP2 features values between 0.44 and 0.47. For the period 2004 to 2007 WaterGAP2 yields better results than for the period 1992 to 2003. This can be explained by the small number of gauging stations during 2004 to 2007, which are dominantly allocated in large river basins. Due to the coarse spatial resolution of WaterGAP2 these basins can better be represented than the average Mediterranean and Black Sea basin. Once again, the reasons for the discrepancy between both models can be found in the different spatial resolution and model structures mentioned in chapter 2.1.

3.2 Sea-mass variation validation

For ocean mass estimates, the amount of continental hydrological leakage which is present in the GRACE basin averages after application of the post-processing filter, depends on the filter smoothing and increases with the radius of the filter. Of course, it also depends on the hydrological model used. The annual amplitudes of the leakage computed from the three global models GLDAS, LAD and WaterGAP2 using the DDK3 filter are significantly different (Table 1). Depending on the model chosen, it ranges between 7 and 23 mm in the Mediterranean Sea and between 24 and 63 mm in the Black Sea. The phase of the annual component is between 41 and 55 degree in the Mediterranean Sea and between 50 and 60 degrees in the Black Sea. The semi-annual component is almost absent in the Mediterranean Sea and is small in the Black Sea.

In Fenoglio-Marc et al. (this issue) we have been using a version of the global model WaterGAP2 with a native resolution of 0.5° to estimate the leakage caused by hydrology of the GRACE signal. We extend here the analysis including the regional model WaterGAP3. In the Mediterranean Sea the annual amplitude of the continental hydrological leakage is smaller when computed from WaterGAP3 (14 +/- 3 mm) than when from WaterGAP2 (19 +/- 3 mm) and peaks in both cases around the 15th February (Table 2, Figure 9). Also in the Black Sea the annual amplitude corresponding to WaterGAP3 is smaller (29 +/- 5 mm) compared to the amplitude corresponding to WaterGAP2 (39 +/- 5 mm) and peaks one week earlier around the 20th of February (Table 3, Figure 9).

The seawater mass (SLV_{mass}) has been obtained by correcting the GRACE/GFZ solution for the continental hydrology leakage computed from WaterGAP3 and WaterGAP2 and by re-scaling as described in section 2.2. The agreement between this GRACE-derived SLV_{mass} (indicated here also with “G-h”) and the SLV_{mass} derived from steric-corrected altimetry (indicated here also with “a-s”) is higher in the Mediterranean Sea than in the Black Sea, independently from the hydrology model used. In both basins the agreement is higher when the WaterGAP2 model is used.

In the Mediterranean Sea correlation and RMS differences of the monthly time-series are 0.83 and 21 mm with WaterGAP3 and 0.86 and 40 mm with WaterGAP2 (Figure 10). The annual signals are comparable and, using WaterGAP2, the GRACE-derived SLV_{mass} (annual amplitude 27 +/- 3 mm peaking on the 18^{th} November) is consistent within the altimetry-derived SLV_{mass}, (annual amplitude 23 +/- 3 mm peaking on the 24^{th} November) within 3 mm and 23 days. With WaterGAP3, the SLV_{mass} has annual amplitude of 35 +/- 4 mm peaking on the 29^{th} December. The higher disagreement (12 mm and 35 days) is due to the smaller amplitude of WaterGAP3. As shown in Fenoglio et al. (this issue) a continental hydrological correction bigger than the correction provided by WaterGAP2 is required when using the MFSTEP steric correction. Figure 11 (top) depicts annual amplitudes and phases for each of parameters entering in the computation of SLV_{mass} and shows the amplitude and phase (h, bleu) of the rescaled hydrology component (34 mm, peaking on the 16^{th} February) that best agrees with the selected altimetry, GRACE data and MFSTEP model.

In the Black Sea correlation and RMS differences of the seawater mass SLV_{mass} monthly time-series are 0.74 and 90 mm with WaterGAP3 and 0.71 and 122 mm with WaterGAP2 (Figure 10). Differences in annual amplitude and phase are smaller (3 mm and 9 days) with WaterGAP2. The altimetry-derived SLV_{mass} (annual amplitude 32 +/- 5 mm peaking on the 20^{nd} April) agrees with the SLV_{mass} from WaterGAP3 and GRACE (annual amplitude 55 +/- 4 mm peaking on April 1^{st}) within 23 mm and 20 days. The lower consistency is due, also in this case, to the smaller amplitude of WaterGAP3, as a higher amplitude of the continental hydrological correction is needed when using the NEMO steric correction (Table 3). Figure 11 (bottom) depicts annual amplitudes and phases for each parameters and shows amplitude and phase (h, blue) of the rescaled hydrology component (74 mm peaking on the 21^{st} February) which best agrees with the given observations from altimetry and GRACE and with the NEMO steric correction.

4. Conclusions and discussion

This interdisciplinary study shows that WaterGAP3 produces improved estimates of total water storage in the Mediterranean and Black Sea watersheds. Concerning the inflow to the Black Sea the Danube provides more than 50% of the total inflow, followed by Don, Dnieper, and Kuban. Generally, the inflow is peaking during spring due to snow melt conditions followed by a smaller rainfall induced peak in late autumn. The inflow to the Mediterranean Sea is dominated by the Po and Rhone rivers which each contribute about 25% of the total inflow, whereas peak flow mostly occurs in winter. Hydrological inflow to the Mediterranean and Black Sea as modeled with WaterGAP3 generally yields good results when compared to

values from literature (Ludwig et al. 2009). Calculated contributions to the total runoff into the Black Sea from individual discharge regions in WaterGAP3 simulations show in general a good agreement to climatological estimates from Jaoshvili (2002). But results reveal a lesser importance of Georgian rivers for the basin's water budget compared to climatology.
The comparison of modeled river runoff of the regional hydrology and water use model WaterGAP3 with its global counterpart WaterGAP2 features advantages for WaterGAP3. When comparing both models to observed river runoff at the 63 stations available for the study region (GRDC 2004) WaterGAP3 results features a coefficient of determination which is about 0.09 higher than those of WaterGAP2 until the year 2000. Thereafter, the number of available station data drops and mainly stations in large river basins remain, which reduces the advantage of WaterGAP3 to 0.04. One of the main reasons for the general discrepancy of both models can be found in the coarse spatial resolution of WaterGAP2. Especially, in the Mediterranean region, small river basins dominate the total inflow into the ocean, which cannot be well represented in the global model WaterGAP2. Also, the comparison between WaterGAP3 and GLDAS shows large advantages for WaterGAP3, whereas it needs to be mentioned that GLDAS data only for two gauging stations in the study region has been available.

We have used WaterGAP3 to compute the continental hydrology leakage correction for GRACE and to derive ocean water mass variation. The comparison of the resulting GRACE-derived mass change with the ocean mass variation derived from steric-corrected altimetry shows a general agreement in both the Mediterranean and Black Sea. Results are however slightly worse as by using WaterGAP2, due to the smaller annual amplitude of the continental leakage corresponding to WaterGAP3. Similar results have been found with GLDAS-CLM (Fenoglio-Marc et al. (this issue)) confirming the similarity between WaterGAP3 and GLDAS. On the other hand, the altimeter-derived estimates depend on the steric correction that is of course not fully perfect. We conclude that the regional model WaterGAP3 is suitable to correct ocean mass changes for hydrological leakage and that both the hydrology and ocean models can benefit from use of GRACE and altimetry in constraining the difference between the water mass derived from the two methods. Small amplitudes and phase differences have been observed and need further investigation.

The regional WaterGAP3 hydrology and water use model needs further developments to include a buffer strip around the current STREMP region to ensure correct application of the GRACE filters, as currently, only the river basins draining into the Mediterranean and Black Sea are being considered. Improvements are planned in modeling the Nile River basin, with alteration of the reservoir algorithm to reflect the maximum annual release of 15 km³ (Nixon 2003) of fresh water into the Mediterranean Sea. Furthermore, it needs to be checked if the water uses from all five modeled sectors represent the current conditions at the downstream area of the Nile River.

We finally expect that the interaction and coupling of the regional hydrology and ocean models using geodetic observations will improve the models. The assimilation of GRACE data in WaterGAP3 as well as the separation of the GRACE signal into different hydrological storage compartments by assimilating remote sensing data will be the next steps. Important improvements to the hydrology model are expected, as GRACE will provide large scale mass constraints to hydrology modeling in regions where observed data are missing.

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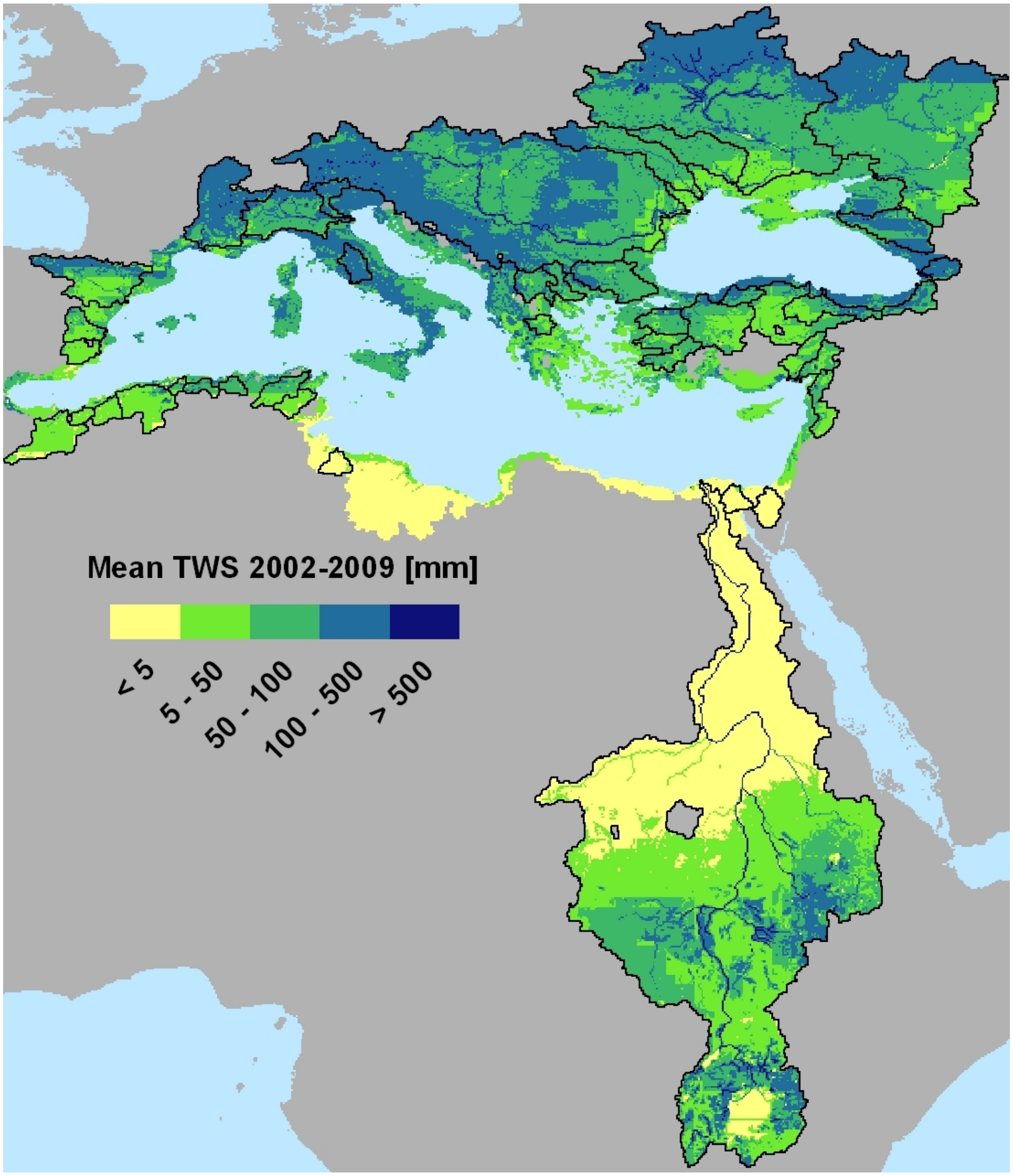


Figure 1: Mean monthly TWS (Total Water Storage) of the Black and Mediterranean Sea river basins as simulated with WaterGAP3. Black borders show the outlines of the 40 largest rivers basins.

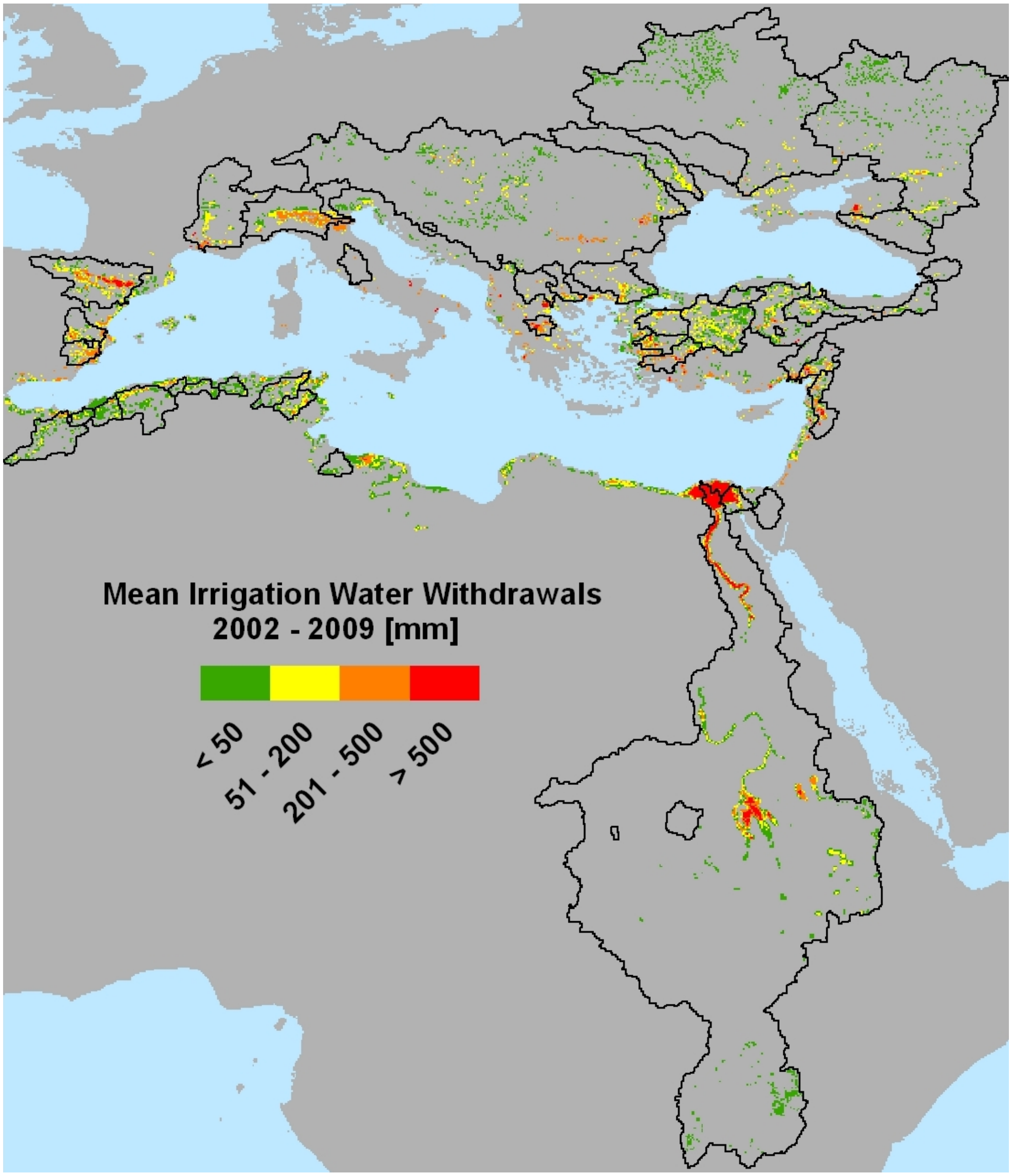


Figure 2: Mean annual irrigation water withdrawals as simulated with WaterGAP3.

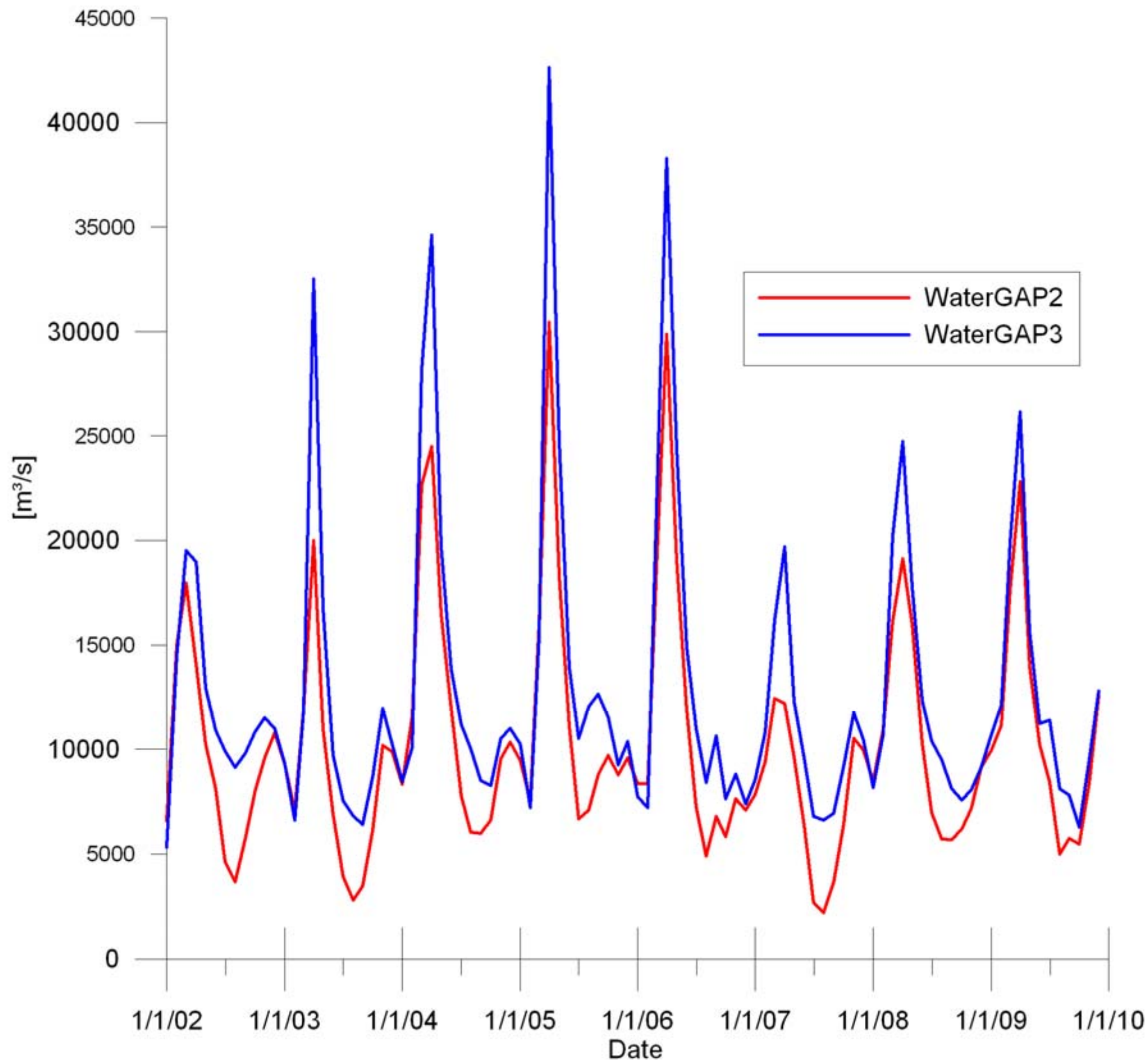


Figure 3: Modeled cumulated monthly inflow into the Black Sea.

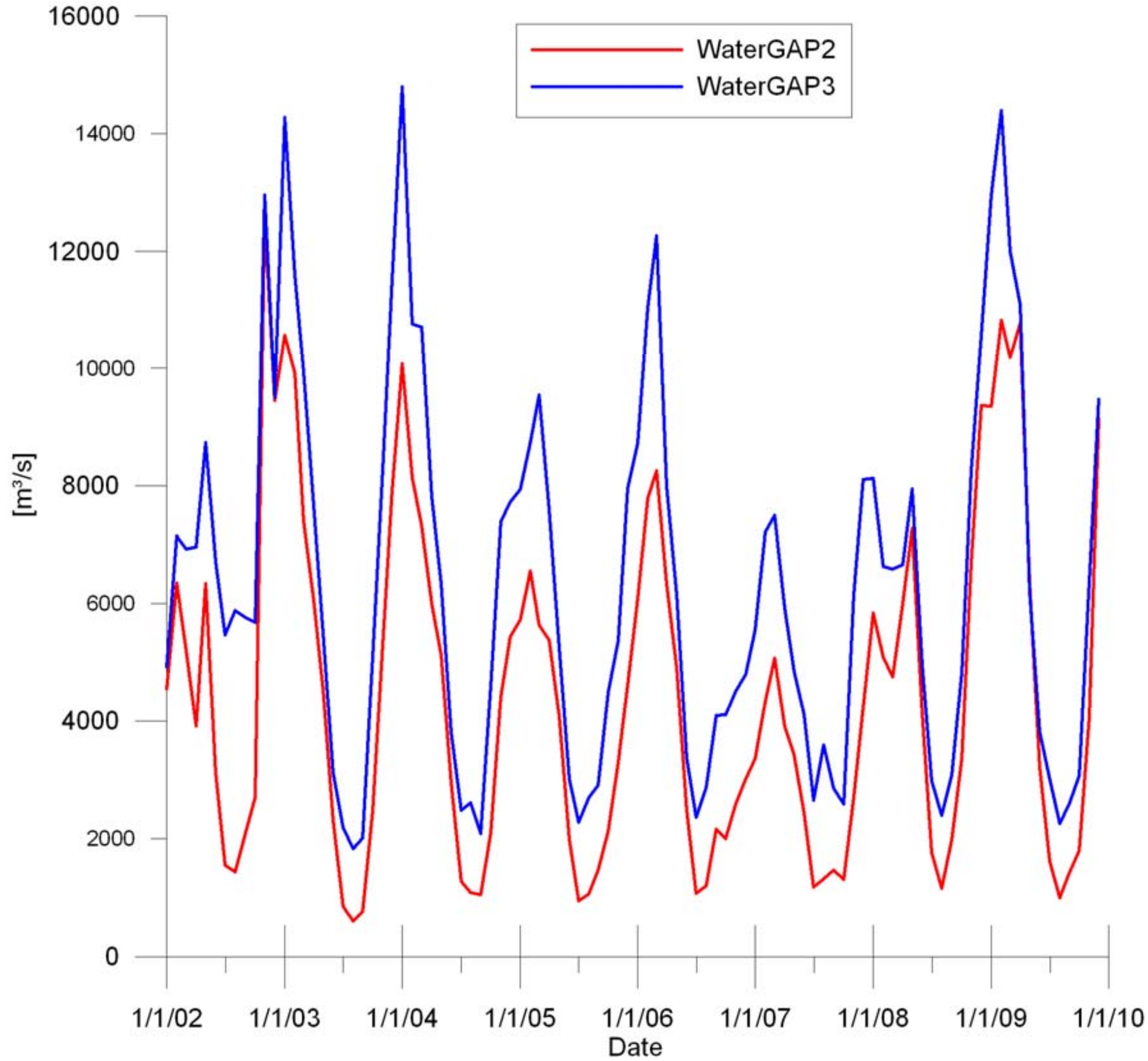


Figure 4: Modeled cumulated monthly inflow into the Mediterranean Sea. Nile runoff data is not included due to its heavy regulation.

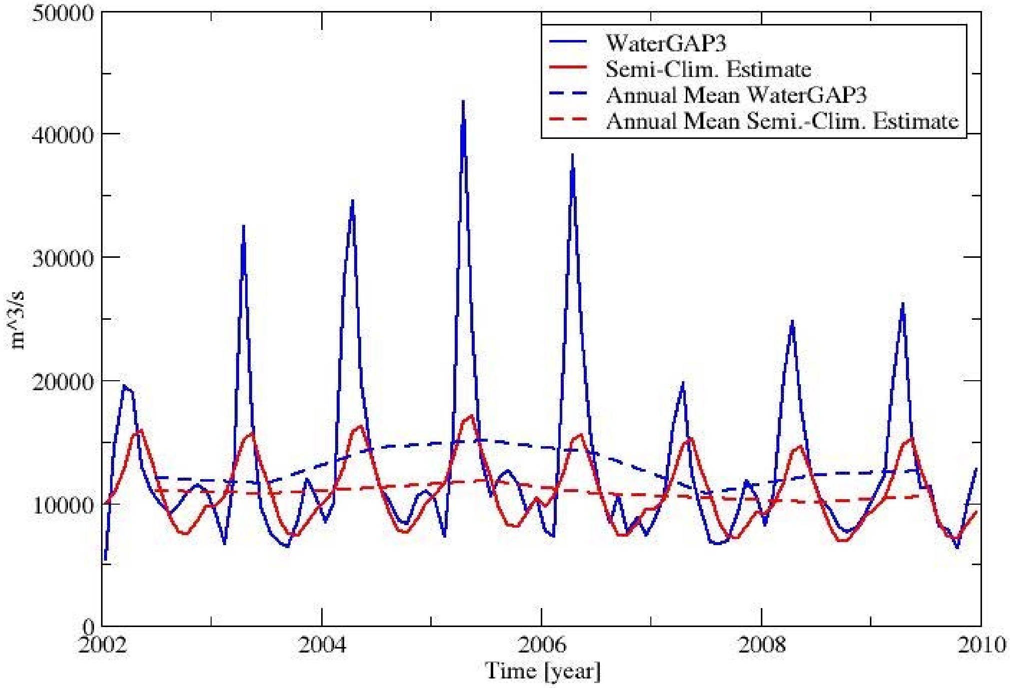
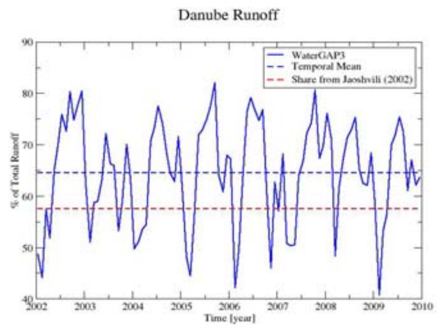
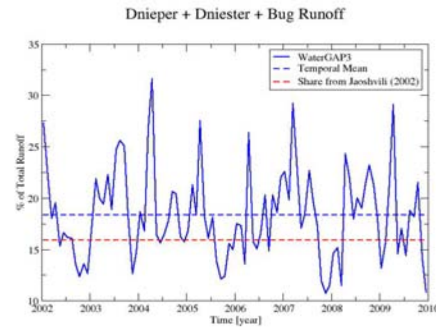


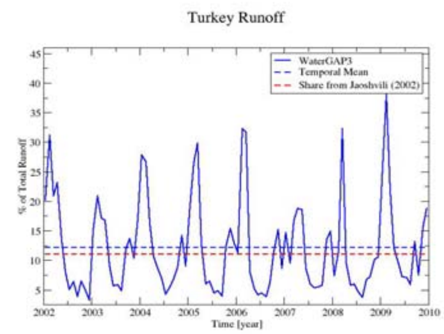
Figure 5: Comparison of total runoff into the Black Sea from semi-climatological estimation from Grayek et al (2010) (red solid line) and from WaterGAP3 (blue solid line). The dashed lines show the corresponding annual mean of the data.



a)



b)



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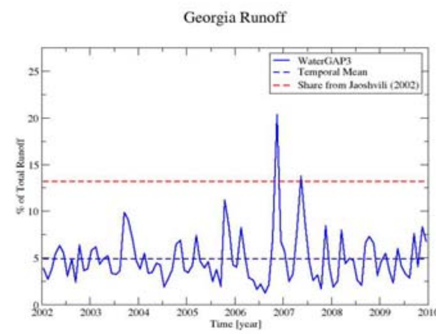


d)

Figure 6: Discharge shares of drainage regions from Jaoshvili (2002) calculated from WaterGAP3 output for the Black Sea (blue line) and the climatological values taken from Jaoshvili (2002) (red line). Russian rivers (1.9% of total runoff in Jaoshvili (2002)) and Bulgarian rivers (0.6% of total runoff in Jaoshvili(2002)) are not illustrated because they are not included in the present WaterGAP3 data.

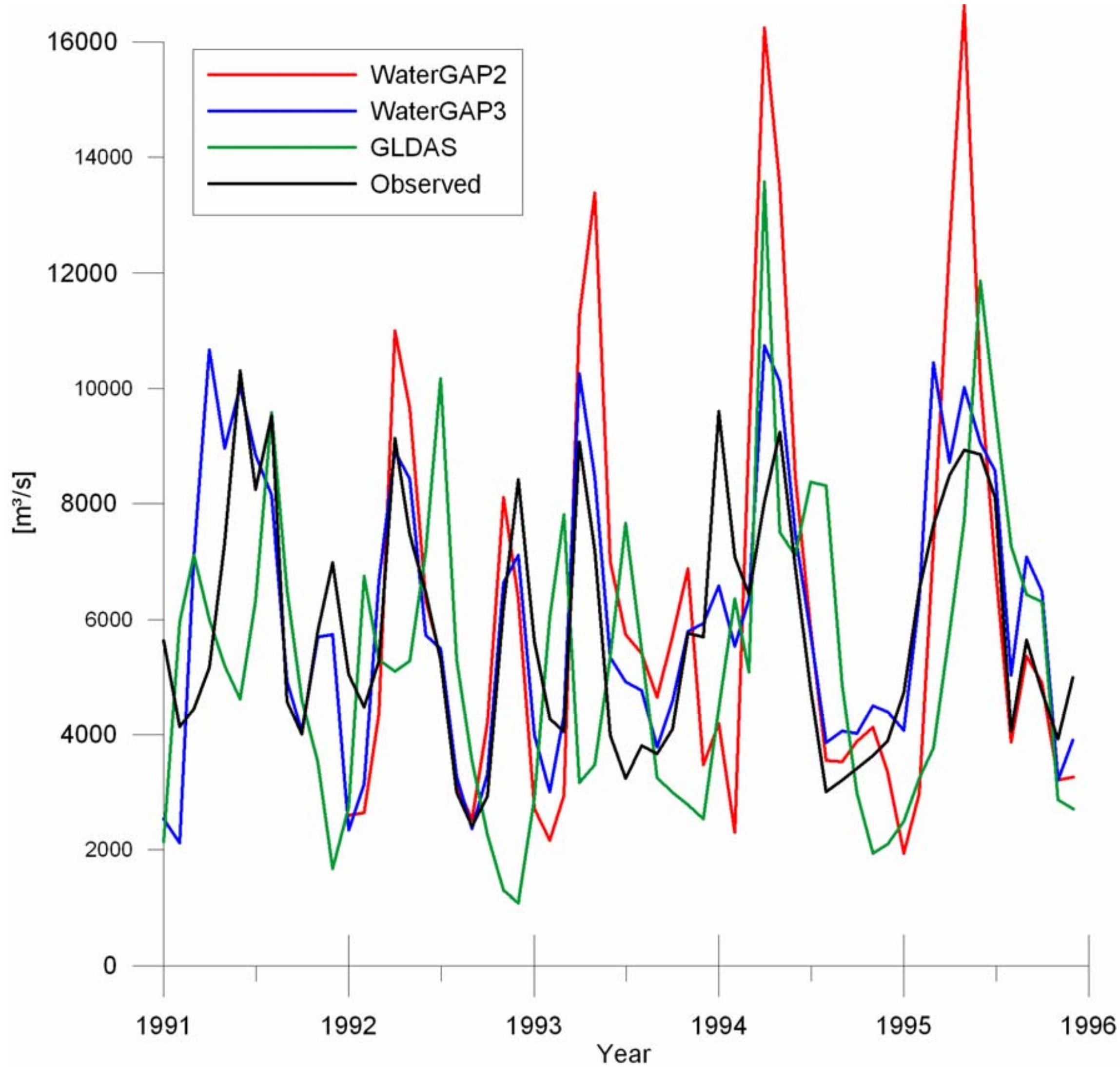
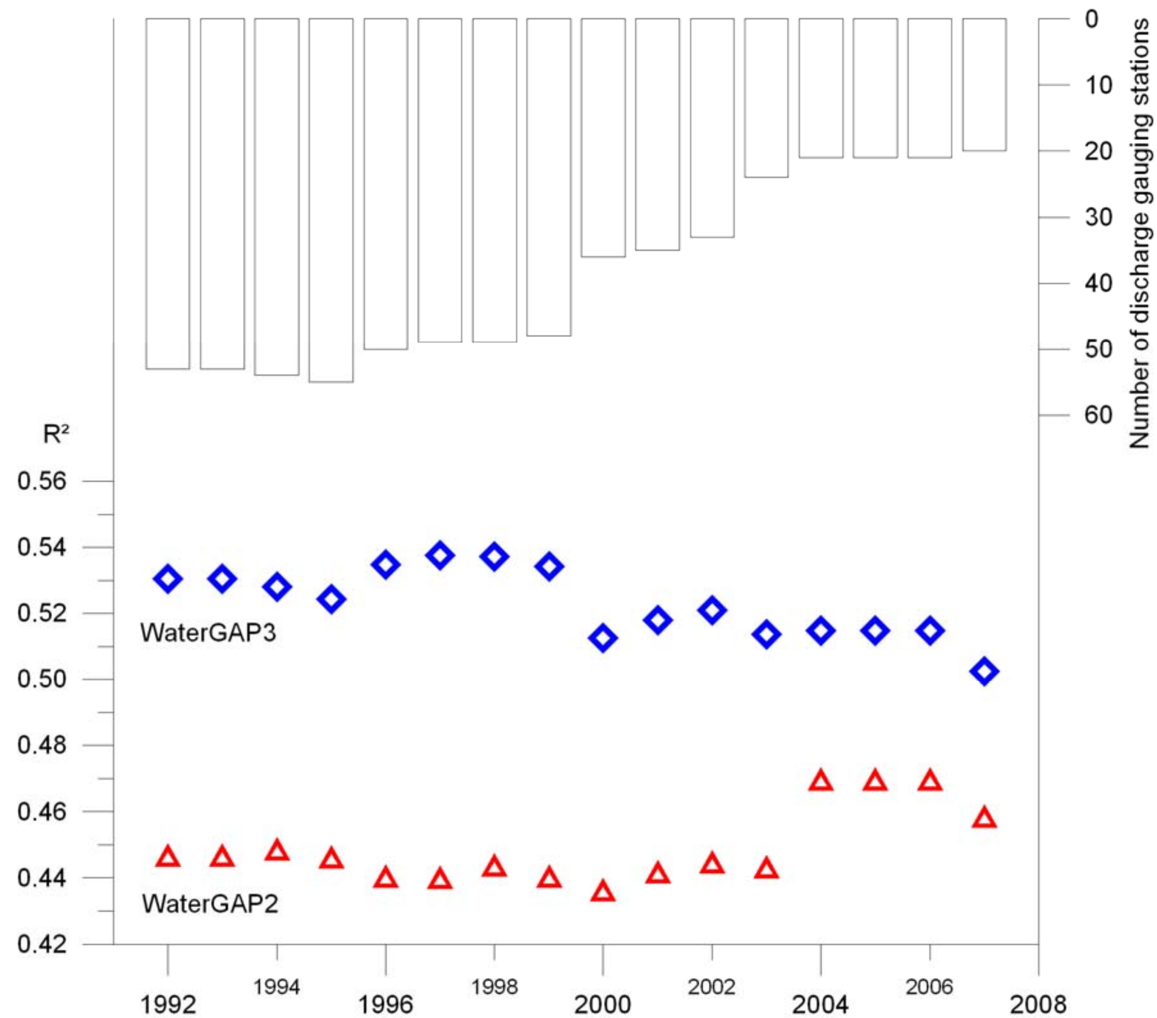


Figure 7: Comparison of modeled river runoff from two different hydrological models and a land surface model (GLDAS) with observed river runoff at the GRDC Station "Ceatal Izmail" (ID 6742900) of the Danube River.



607
608 Figure 8: Evaluation of the goodness of modeled river runoff (compared to observed data)
609 with regard to number of discharge gauging stations available.

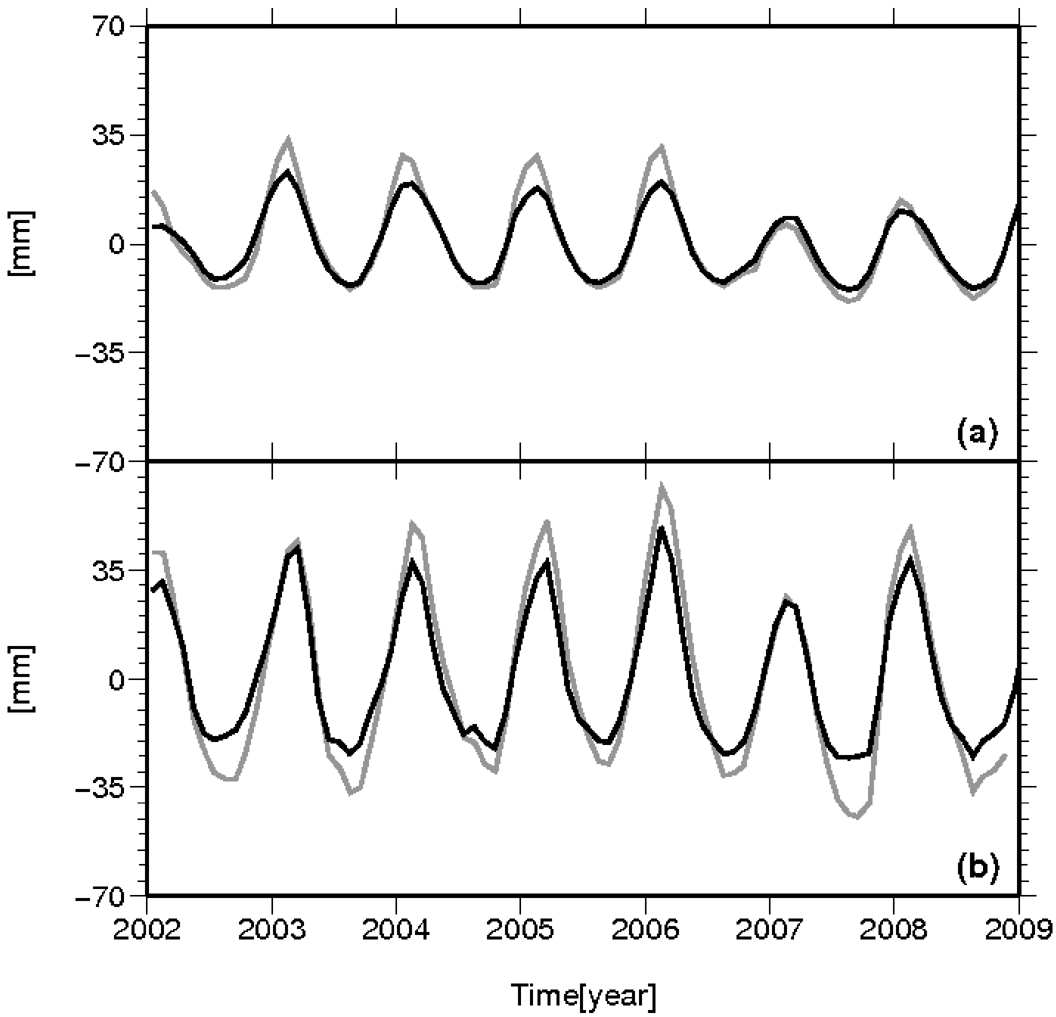


Figure 9: Hydrological leakage included in the seawater change estimated from GRACE in Mediterranean Sea (a) and in Black Sea (b) derived using WaterGAP2 (grey) and WaterGAP3 (black). Re-scaling is not applied.

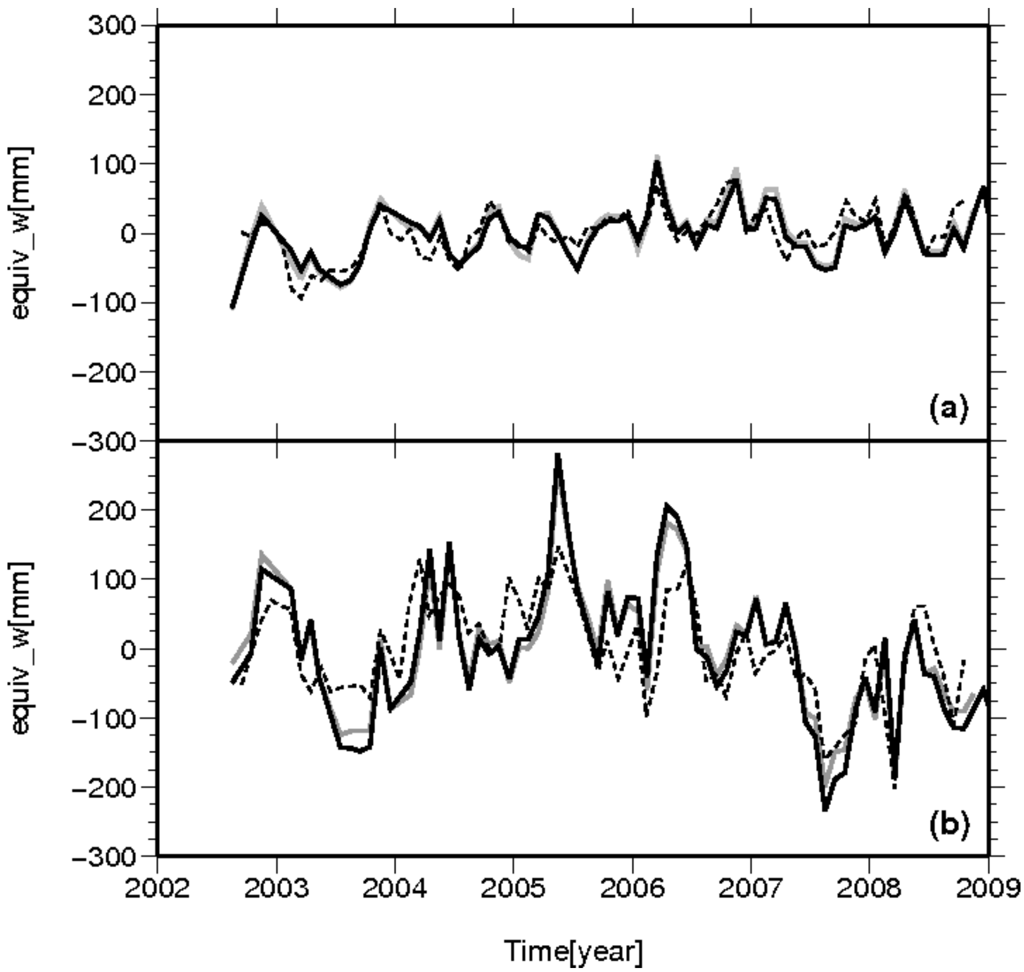


Figure 10: Basin average in Mediterranean Sea (a) and in Black Sea (b) of mass-induced sea level derived from GRACE/GFZ solutions corrected for the hydrology leakage using WaterGAP3 (black) and WaterGAP2 (grey). Mass-induced sea level derived from steric-corrected altimetry (dashed line) is also shown.

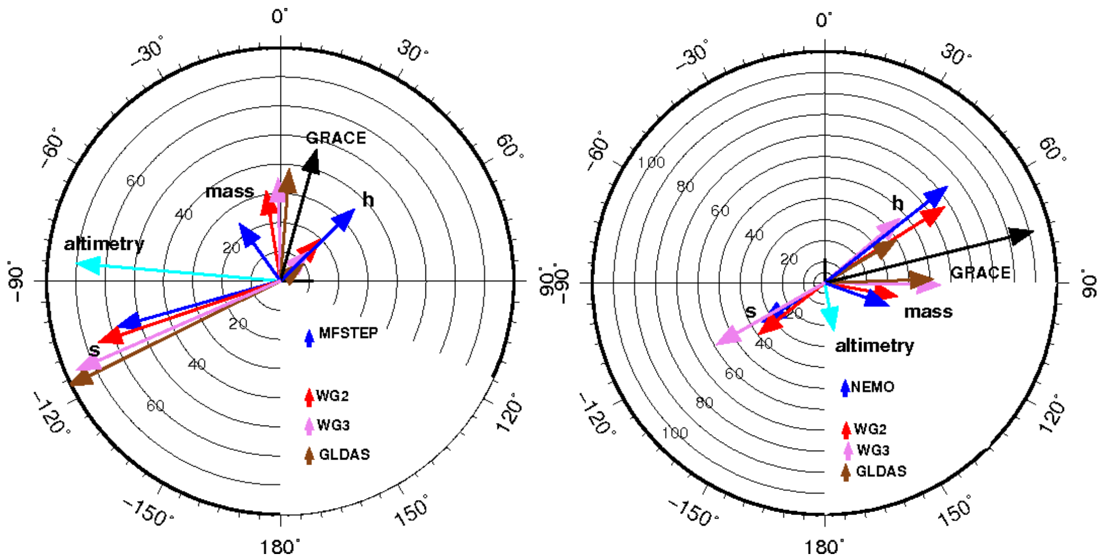


Figure 11: Annual amplitudes and phases in Mediterranean Sea (left) and Black Sea (right) of observed and derived parameters (water mass, steric correction (s) and continental hydrology correction (h)), for selected hydrology and ocean models. GRACE and altimetry basin averages are kept fixed. In each basin a regional ocean model and three hydrology models are used. Re-scaling has been applied.

Table 1: Annual (*A*) and semi-annual (*SA*) amplitude (*Amp*, mm) and phase (*Ph*, days) of monthly averages of hydrology leakage from different hydrology models due to filtering (DDK3) and to truncation (degree 100) in the Mediterranean (*MED*) and Black Sea (*BS*).

Model	MED A Amp	MED A Ph	MED SA Amp	MED SA Ph	BS A Amp	BS A Ph	BS SA Amp	BS SA Ph
GLDAS-CLM01	7 +/- 3	55 +/ 22	1 +/- 3	171 +/- 4	24 +/- 3	60 +/- 7	1 +/- 3	71 +/- 58
LAD	23 +/- 3	41 +/ - 8	2 +/- 3	34 +/- 2	63 +/- 3	58 +/- 3	4 +/- 3	74 +/- 24
WaterGAP2	19 +/- 3	45 +/- 8	4 +/ 3	26 +/- 2	39 +/- 3	57 +/- 4	5 +/- 4	47 +/- 12
WaterGAP3	14 +/- 3	44 +/- 11	2 +/- 3	34 +/-20	29 +/- 5	50 +/- 6	6 +/- 3	53 +/- 13

Table 2: Annual (A) and semi-annual (SA) amplitudes (*Amp*, mm), phases (*Ph*, days) and trends of monthly basin averages of water mass variation SLV_{mass} from steric-corrected altimetry (*a-s*) and hydrology-corrected GRACE (*G-h*) in the Mediterranean Sea from WaterGAP2 (*WG2*) and WaterGAP3 (*WG3*), continental hydrology leakage correction simulated by the two models (h_{WG2}, h_{WG3}) and derived from comparison of GRACE with steric-corrected altimetry, steric correction simulated by an ocean model and derived from comparison of altimetry with hydrology-corrected GRACE. The re-scaling factor 1.4 has been applied.

	A Amp (mm)	A Ph (days)	SA Amp (mm)	SA Ph (days)	Trend (mm/yr)
SLV_{mass} = a-s_{MFS}	24 +/-3	329 +/- 6	13 +/- 3	119 +/- 3	8.3 +/-3
SLV_{mass} = Scaled (G-h_{WG2})	27 +/- 4	352 +/- 3	19 +/-5	123 +/- 5	5.3 +/- 2
SLV_{mass} = Scaled (G-h_{WG3})	35 +/- 4	364 +/- 3	17 +/- 4	125 +/- 3	5.5 +/- 2
h_{WG2}	20 +/- 3	44 +/- 6	2 +/- 3	34 +/- 3	-1.0 +/- 0.6
h_{WG3}	14 +/- 3	44 +/- 6	2 +/- 3	34 +/- 6	-1.2 +/- 6
h = Scaled (G – a + s_{MFS})	34 +/- 3	47 +/- 7	2 +/- 5	160 +/- 3	-8.7 +/- 2.1
s_{MFS}	58 +/- 4	258 +/-4	1 +/- 4	86 +/-4	-10.1 +/- 0.6
s_{wg2} = Scaled (a-G+h_{WG2})	66 +/- 4	255 +/- 4	6 +/-3	31 +/- 3	-5.3 +/- 1.1
s_{wg3} = Scaled (a-G+h_{WG3})	76 +/- 4	250 +/- 4	5 +/- 4	50 +/- 4	-5.8 +/- 1.3

Table 3: as Table 2 for the Black Sea. The re-scaling factor 1.7 has been applied.

	A Amp (mm)	A Ph (day)	SA Amp (mm)	SA Ph (day)	Trend (mm/yr)
$SLV_{mass} = a\text{-}s_{NEMO}$	32 +/-5	111 +/- 10	33 +/- 3	163 +/- 10	-12.2 +/- 2
SLV_{mass} = Scaled ($G\text{-}h_{WG2}$)	35 +/- 4	102 +/- 3	45 +/-8	147 +/- 10	-12.2 +/- 2
SLV_{mass} = Scaled ($G\text{-}h_{WG3}$)	55 +/- 4	92 +/- 3	49 +/- 4	148 +/- 3	-13.2 +/- 2
h_{WG2}	68 +/- 7	58 +/- 6	10 +/- 5	50 +/- 5	-0.3 +/- 2
h_{WG3}	47 +/- 3	50 +/- 6	11 +/- 3	53 +/- 6	-1.2 +/- 2
h = Scaled ($G – a + s_{NEMO}$)	74 +/- 3	52 +/- 7	22 +/- 5	111 +/- 3	-2.5 +/- 2.1
s_{NEMO}	35 +/- 4	241 +/-4	3.5 +/- 4	38 +/-4	-0.3 +/- 0.6
s_{wg2} = Scaled ($a\text{-}G+h_{WG2}$)	40 +/- 4	235 +/- 4	19 +/-3	32 +/- 3	0.1 +/- 1.1
s_{wg3} = Scaled ($a\text{-}G+h_{WG3}$)	59 +/- 4	243 +/- 4	20 +/- 4	35 +/- 4	0.5 +/- 1.3