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TERRESTRIAL WATER STORAGE CHANGE OF EUROPEAN RUSSIA AND ITS IMPACT ON WATER BALANCE

ABSTRACT. Terrestrial water storage has a significant impact on the water balance of river basins. The analysis of its changes in the European part of Russia (EPR) using the GRACE (Gravity Recovery and Climate Experiment) data showed that its reduction was approximately 150 mm for 2002-2015 for the south of EPR, especially the Don basin, which is caused rather by a decline in the storages of surface and ground waters then to changes in soil waters. Quasilinear relation between the values of terrestrial water storages and a river runoff for the period of a summer low water level for a number of rivers has been revealed.

KEY WORDS: GRACE, water balance, European Russia, soil water content.

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INTRODUCTION

In 1896 A. Penk proposed the equation of the water balance of a river basin for long-term period. E. V. Oppokov in 1904 proposed the equation of the water balance of a river basin for some period as the sum of precipitation (*P*), evaporation (*E*), a river and underground runoff (*R*), and a terrestrial water storage change (*TWSC*) equal to zero in its most general form (Babkin and Vuglinsky 1982)

$$P_t - E_t - R_t \pm TWSC_t = 0 \qquad (1)$$

where *t* indicates some time interval. Further in the text this index is omitted.

The total terrestrial water storages (*TWS*) are the storages of surface and ground waters in all aggregate states. As a rule, individual *TWS* components are of interest. Thus, the snow water equivalent (*SWE*) is one of the key predictors of spring high water runoff and the maximum values of a water flow for the rivers with a cold climate. The risk of hydrological drought depends on groundwater storages. The soil water content (*SWC*) largely determines the nature of the interaction of the atmosphere with the underlying surface (Kumar et al. 2016), and is also a factor in the formation of the maximum runoff. Terrestrial water storage change TWSC is the difference between end-of-period and beginning-of-period *TWS*.

There are several methods to determine the value of *TWSC*. One of them used in the GRACE (Gravity Recovery and Climate Experiment) project is based on the impact of water mass redistribution on the Earth's gravitational field. The basis of the GRACE system is 2 satellites equipped with microwave range finders, star cameras, accelerometers and GPS signal receivers (Zotov et al. 2015). A component due to a change in *TWS* is obtained using the models of circulation of the atmosphere and the ocean, as well as the dynamics of the earth's crust from the Earth's gravitational field. Thus, the obtained data on a *TWS* change are a solution to an inverse regularization problem. Since the solution to this problem is unstable, the final result depends on a method for solving the problem. There are two main approaches to solving this problem. The first is in the form of Stokes coefficients, global, the solution is sought for the entire land. The second is in the form of mascons, local, the solution is sought for each local area, for example 3° hexagons (Save et al. 2016). The result of the calculation presents an anomaly of terrestrial water storage (not its absolute value) TWSA, for any period, most often in one month.

GRACE data are used in meteorology to validate a reanalysis (Springer et al. 2017) and improve a surface air temperature forecast (Lin et al. 2016), and in glaciology to validate and calibrate the melting of a seasonal snow cover and glaciers (Wahr et al. 2016; Chen et al. 2017a; Schlegel et al. 2016). The use of GRACE data in hydrology is mainly limited to the issues of water runoff estimation, although there is an example of their use for studying the dynamics of river sediments (Liu et al. 2016). The study of a river runoff using GRACE is concentrated in two main trends: 1) the calculation of various components of water balance using remote sensing data (RSD) and a weather reanalysis using the water balance equation; 2) the assimilation of GRACE data into hydrological models and LSM (land surface model). The first trend appeared first of all. An unknown member of water balance (usually a river runoff) is calculated therein from the known values of the remaining terms. Despite the low accuracy of such calculations in comparison with the measurements or simulation results, they make it possible to estimate it for large areas there are few measurement data for (Li et al. 2016; Lorenz et al. 2014) or they are not publicly available (North Korea (Seo and Lee 2017)). Due to the rough spatial resolution of GRACE data, the calculation of a river runoff directly from the changes in channel storages, without using other RSD, is only possible on the Amazon River, the largest river (Eom et al. 2017). In a number of cases, such as the dynamics of the level of the Caspian Sea (Chen et al. 2017b), the world ocean (Chambers et al. 2017), the water resources of arid lands (Deng and Chen 2017; Forootan et al. 2017, Frolova et al. 2017) and wetlands (Xie et al. 2016), *TWSC* is precisely of interest. If the dynamics of water storages varies significantly in different reservoirs in a basin, it is possible to split *TWSC* into separate components using a spectral analysis (Andrew et al. 2017).

The first study within the framework of the second trend is (Zaitchik et al. 2008), where the GRACE data were used to improve the reproducibility of the flood in the Mississippi basin in June, 2008. The assimilation of measurement data into a model is based on the Bayesian approach, when the measurement data are used to find the posterior probability distribution function (PDF) of the model parameters. In the case of a nonlinear system, which is a hydrological cycle, there is no analytical solution, and a numerical solution is required. A group of variational methods and statistical sequential analysis methods can be distinguished from them (Khaki et al. 2017). The former are little used due to their laboriousness. The latter are mainly an ensemble Kalman filter, both a deterministic and stochastic one, an ensemble Kalman smoother, a particle filter and a non-Gaussian rank histogram filter (Khaki et al. 2017). Together with the GRACE data, other RSDs, such as SMOS (soil water content) and MODIS (snow cover area) can be assimilated into a model. The experience of GRACE data assimilation showed that the decrease in the error in reproducing ground water storages and the soil water content in a layer of several dozen centimeters was the greatest (Tian et al. 2017; Khaki et al. 2017; Kumar et al. 2016; Tangdamrongsub et al. 2017) - by 15-30%. Especially important is the assimilation of GRACE data when there is a change in basin water storages as a result of anthropogenic activity (pumping out) that is not taken into account in a model, although this is what may lead to the appearance of pseudotrends for other parameters (Girotto et al. 2017). The use of GRACE can also improve the reproducibility of snow water equivalents, mainly in high latitudes and highland areas (Tibet). This is due to a relatively low quality of the input meteorological information in these areas (a sparse observation network and systematic errors) that is often the main source of model errors. In addition, GRACE errors do not depend on such factors as the forest coverage, height and structure of the snow cover that are critical for a microwave survey (Zhang and Yang 2016; Lin et al. 2016; Forman et al. 2012). For water flow rates and evaporation, the change in the accuracy of calculation, when using GRACE, is generally insignificant, and can give both a positive and negative effect.

The question of an impact of total of water storages on the remaining components of water balance (*P, E, R*) has been considered for a long time. The relation between modulus of flow Q and *TWS* may be written as (Klemes 1974; Klemes 1978).

$$Q = a(TWS)^c \qquad (2)$$

where a and c are coefficients, Q – modulus of flow ($(m^3/s)/km^2$).The power dependence for some assumptions about the similarity of a river network can be obtained from a kinematic wave equation for a slope runoff (Dolgonosov 2008), moreover, the exponent can be in the range from 1.5 to 3 depending on the preferred type of a runoff (1.5 for a turbulent and 3 for a laminar one). The value of the autocorrelation coefficient, the dispersion of a river runoff and the correlation coefficient between precipitation and a river runoff depend on the dependence $R = f(TWS)$ (Frolov 2011; Frolov 2014).

The paper aims at analyzing a change in TWS in the European part of Russia (EPR) for 2002-2015 and studying the relation between terrestrial water storages and a river runoff.

MATERIALS AND METHODS

The source of the GRACE data was the site (JPL GRACE data) that contains the processing materials of a geophysical institute (GFZ, Potsdam, Germany), the Center for Space Research (CSR, Austin, the USA) and the Jet Propulsion Laboratory (JPL, Pasadena, the USA). The values of terrestrial water storages were calculated as the average values between the blocks of these three centers. As it is shown, the arithmetic mean between these 3 archives has the smallest error compared to each of the archives individually (Sakumura et al. 2014). The data has a resolution of 1° in latitude and longitude and 1 month in time (monthly averages). There are gaps in the GRACE data series, which is due to the deterioration of batteries on satellites and the ability to maintain the required voltage only for certain orbital parameters (https://grace.jpl.nasa.gov). Thus, there are 19 gaps in the series from April 2002 to January 2017, i.e. the actual length of a series is 159 months. 13 passes refer to the period after 2010. For the same reason, the measurements for some months have a reduced accuracy. The lowered accuracy is also characteristic for the data for 2002, which is due to the incompletely adjusted operation of all devices in the first months of the operation of satellites. Due to the presence of gaps in the data, the period of 2002-2015 was used for calculating the minimum annual water storages, and the period of 2003-2015 was used for the maximum annual water storages.

Materials on the regime snow-measuring surveys for 284 routes on EPR for 1966-2015 were taken from the site meteo.ru. The study also used the data of a ERA-Interim reanalysis (Albergel et al. 2012) on a soil water content. The expedited data for 0, 6, 12 and 18 hours were used to calculate the average monthly values (at the levels of 0-7 cm, 7-28 cm, 28-100 cm and 100-289 cm). ERA-Interim covers the period from 1979 to 2017 with a resolution of about 0.75° in latitude and longitude. The data for 31 gauging stations were used to calculate the average monthly runoff (Table. 1)

Table 1. Parameters from the power function (2), linear function (3) and their accuracy (4) in each of the 31 river basins using TWSA to calculate river discharge

river basin	area, th. km^2	power function				linear function			
		D, %	err_{rel}, %	a	c	D, %	err_{rel}, %	$a_{lin}\cdot 10^9$, 1/s	$b\cdot 10^3$, $m^3/(s*km^2)$
Samara -Kargala	29.6	73.5	21.2	$5.7*10^{-11}$	2.93	68.1	25.6	12.1	-2.49
N. Dvina-Abramkovo	220	63.9	13.9	$5.57*10^{-8}$	2.06	65.0	13.7	44.9	-6.32
Mezen - Bolshaya Pyssa	16.1	41.3	27.5	$2*10^{-9}$	2.77	41.2	26.7	104	-17.0
Pechora-Troitsko-Pechorsk	35.6	34.1	23.0	$9.07*10^{-6}$	1.33	35.0	22.6	81	-6.36
N. Dvina -Ust Pinega	350	75.3	13.7	$5.06*10^{-9}$	2.51	75.9	13.0	59.3	-9.54
Pechora - Oksino	310	47.7	15.5	$1.05*10^{-5}$	1.32	48.3	15.3	87.3	-6.19
Mezen - Malonikolskaya	56.4	55.0	22.7	$4.95*10^{-10}$	3.00	53.4	23.3	102	-17.5
Pechora - Ust -Tsilma	250	41.7	19.0	$1.19*10^{-5}$	1.27	42.2	18.7	71.7	-4.64
Neva-Novosaratovka	280	24.8	11.6	$2.99*10^{-5}$	1.00	31.2	11.0	20.6	2.99
Volga - Staritsa	21.1	38.9	33.2	$1.18*10^{-8}$	2.24	39.4	34.8	37	-6.91
Oka - Kaluga	54.9	68.6	10.7	$1.03*10^{-6}$	1.39	67.4	10.9	12.9	-0.964
Oka - Polovskoe	99	68.2	13.4	$1.14*10^{-6}$	1.41	66.7	13.8	15.6	-1.11
Oka - Murom	190	81.3	9.3	$2.12*10^{-6}$	1.27	80.4	9.5	11.9	-0.641
Oka-Gorbatov	240	68.5	10.9	$1.53*10^{-6}$	1.33	67.5	11.0	13.4	-0.877
Moksha-Shevelevsky Maidan	28.6	59.8	11.1	$5.21*10^{-6}$	1.00	59.8	10.9	5.2	0.055
Kama-Gayna	27.4	55.3	19.2	$2.36*10^{-8}$	2.19	55.4	19.8	43.2	-6.62
Kama-Bandug	46.3	61.4	28.3	$2.09*10^{-10}$	3.00	56.6	31.0	51.8	-9.78
Vyatka - Rabino	30.9	39.4	21.9	$3.47*10^{-6}$	1.48	39.2	22.2	75.7	-6.90
Chusovaya-Lyamino	21.5	63.6	31.4	$2.31*10^{-10}$	3.00	53.1	37.0	68.1	-14.0
Belaya-Sterlitamak	21	44.2	20.6	$4.82*10^{-7}$	1.47	42.5	20.8	9.75	-0.786
Belaya Ufa	100	38.4	18.9	$7.7*10^{-6}$	1.07	38.4	18.9	11.9	-0.164
Belaya -Birsk	120	50.9	17.1	$2.43*10^{-6}$	1.26	50.3	17.3	13.4	-0.742
Vyatka-Ustyevskaya	16.5	37.5	34.0	$1.84*10^{-10}$	3.00	32.4	35.9	41.8	-7.38
Vyatka - Kirov	48.3	55.3	20.8	$6.42*10^{-10}$	2.75	51.5	22.7	34.3	-5.90
Vyatka-Kotelniki	72	56.3	15.5	$2.82*10^{-8}$	2.07	55.0	16.1	24.5	-3.51
Vyatka-Arkul'	96.9	50.8	20.7	$3.18*10^{-8}$	2.05	49.8	21.2	25	-3.62
Vyatka - Vyatskie Polyany	120	50.5	15.6	$3.3*10^{-7}$	1.66	48.8	15.9	22.1	-2.28
Don-Zadonsk	31.1	44.8	13.0	$9.6*10^{-6}$	1.00	53.3	11.9	6.89	0.983
Don - Liski	69.5	78.9	8.8	$6.3*10^{-6}$	1.00	78.9	8.9	6.25	0.021
Don-Kazanskaya	100	82.0	10.5	$5.92*10^{-7}$	1.39	80.8	10.9	7.74	-0.652
Don-Belyaevskaya	200	89.0	7.4	$1.03*10^{-6}$	1.27	88.6	7.7	6	-0.382

Revealing a relation between *TWS* and the modulus of flow *(Q)* using the GRACE data in the form of a power dependence that has a physical basis (Dolgonosov 2008; Frolov 2011, 2014) is complicated by the fact that it is not the absolute amount of water in the basin that the available TWS values express, but its anomaly *(TWSA)*, relative to the average value for any period taken as zero. As a result, the TWSA values are negative for some months. To avoid this, the minimum value for the entire observation period was subtracted and 200 mm was added for each basin from the *TWSA* series. To construct the dependences, we used both a power function (2), and a linear one

$$Q = a_{lin} TWSA + b \tag{3}$$

Where alin is a coefficient (1/1000 s). The use of a linear function instead of a power one can be permissible with sufficiently small changes in TWS, due to the differentiation of the latter (Frolov 2011). In addition, application of a linear function allows the use of the water storage anomalies instead of their absolute values, because the dependences obtained will differ only by a constant. The smoothed average monthly values of *Q* and *TWSA* (using the moving average method with a window width of two months) from July to October were used as the data to construct the dependence. In order to evaluate the accuracy of the dependences obtained, we used such parameters as the determination coefficient (*D*) and the mean absolute relative error (err_{rel}) expressed as

$$err_{rel} = \sum_{i=1}^{n} \left| (Q_{obs,i} - Q_{cal,i}) * 100 / Q_{cal,i} \right| / n \tag{4}$$

where $Q_{obs,i}$ is a measured modulus of flow value for *i* month and $Q_{obs,i}$ is a calculated modulus of flow for *i* month.

RESULTS AND DISCUSSION

In the second half of the 20th century, there was an increase in terrestrial water storages in EPR, both due to the growth of *SWC*, and the rise of the groundwater level. On average for EPR, the growth of *SWC* can be estimated as 50 mm in the last decade of September. The data of water balance stations indicate an increase in the level of ground waters over the period of 1950-1990 (Water Resources of … 2008). The main growth was in the mid-1970s and it was 50-130 cm by the early 1990s. Since the average active porosity in EPR is about 15% in the first hydrodynamic zone (with a capacity of about 100 m in EPR), it is possible to estimate approximately a change in groundwater storages by 100 mm. However, a decrease in water storages could be in deeper aquifers, especially where depression pits are formed, the largest of which are in the Central Federal District (Dzhamalov et al. 2015). On the other hand, in the second half of the 20th century, a lot of water storages were created in EPR, which not only increased the total water storages of the area by themselves, but also led to a local rise in the groundwater level. The maximum snow water equivalent changed insignificantly for the period of 1966-2002, having grown for the field routes and decreasing for the forest ones.

The GRACE data indicate a change in *TWSA* in EPR for 2002-2015 already. The months of the minimum ($TWSA_{min}$) and the maximum ($TWSA_{max}$) annual water storages were calculated for this period to analyze the seasonal movement of *TWSA*. Clear zoning is noted both for the time of the maximum and the minimum water storages anomaly. The earliest TWSAmin are formed in the far north - in August (and in July in some places). For the Upper Volga and Kama basins, $TWSA_{min}$ can already be noted in September. South of 53°, TWSAmin are noted in October. It is characteristic up to 50° N that as moving from the north-east to the south-west of the area, the time of $TWSA_{max}$ formation shifts from later dates (May) to earlier ones (March). The fact that $TWSA_{max}$ are later again south of 50° is due to lack of a sufficiently strong stable snow cover in the region (on the lowland and in the foothills), which would determine the peak of the maximum water storages as the peak of the maximum water storages in the snow.

$TWSA_{min}$ (the minimum monthly average value of *TWSA* for the year) hardly changed in the north for the period of 2002-2015 (Fig. 1a). The growth of more than 5 mm/year is only noted in the northwest (in places)

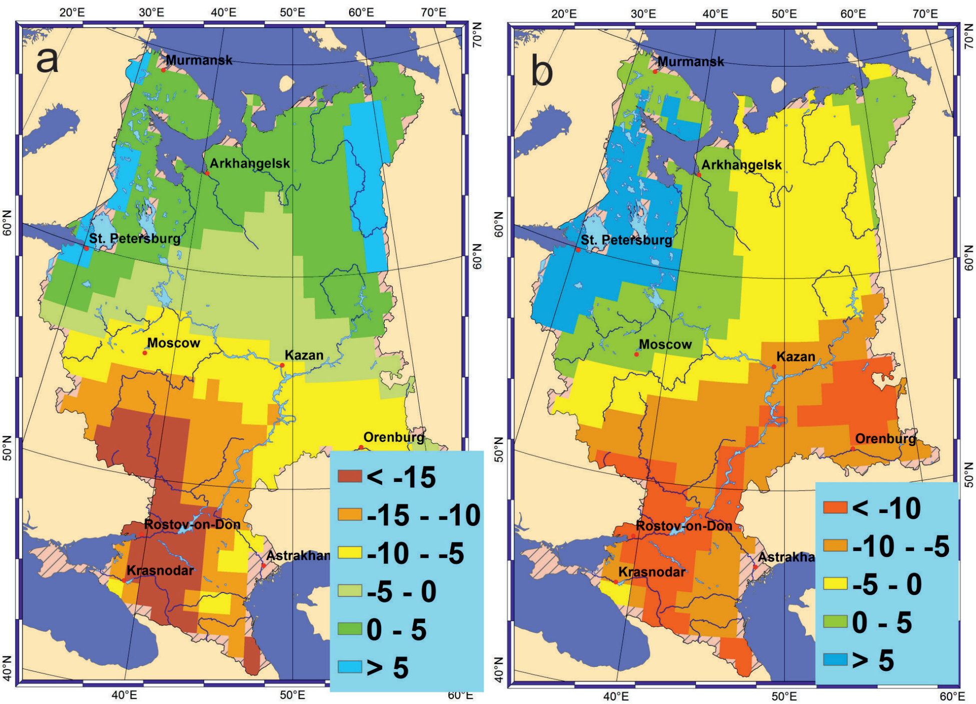

Fig. 1. Linear trend coefficient of the minimum (a) and maximum (b) water storages anomaly for 2002-2015, mm/year (according to GRACE)

and in the northeast (mainly in the Pechora basin). There was a slight decrease in water storages, within 10 mm/year, for the basins of the Kama, the Ural, the Upper Volga and especially the Oka River. The most significant decrease was in the Don basin (to a lesser extent, in the Khoper and Medveditsa basin) and the Kuban basin, where linear trend coefficient or average rate of change it exceeded 15 mm/year.

The change in $TWSA_{max}$ was not so significant - the positive trends do not exceed 10 mm/year and the negative ones - 15 mm/year (Fig. 1b). An increase in $TWSA_{max}$ is noted practically for the whole northwest of EPR. The areas of a decrease in the maximum water storages, as well as the minimum ones, are located in the south of EPR. At the same time, for $TWSA_{max}$, in contrast to $TWSA_{min}$, the decrease in the Kama and Ural basins (more than 10 mm/year) is comparable to that of the Don and Kuban basins.

In general, there were not any *TWSA* changes over the period from 2002 to 2015, although the years with the minimum *TWSA* values (2010, 2011 and 2014) refer to the second half of the period. In the Don basin, which underwent the most significant changes in *TWSA*, there was a decrease in *TWSA* in 2007-2010. Further on, the decline almost ceased, but the growth did not change either. The lowest water storages in the Don basin were in 2015, when their $TWSA_{max}$ was at a level of $TWSA_{min}$ for 2002-2007. The situation in the Oka basin is similar to that in the Don basin. However, the value of *TWSA* has started to grow again there since 2010, having reached its maximum in 2013, when the water storages in the basin were at a level of 2002-2007. But in 2014 and 2015 there was a sharp decrease again. The minimum values of $TWSA_{max}$ were in 2015. The changes in the northern catchments areas were smaller, and they were not unidirectional. A period of a decline in 2002-2006, of a rise in 2006-2007, then a decline from 2007 to 2012 again and a rise to 2016 can be noted for the Pechora basin. The situation with the Northern Dvina River is the same. A growth in 2002-2004 has been noted in the basin of the Neva River (the area of the Ladoga and Onega lakes was not taken into account when calculating

TWSA). There were no directed multi-year changes for the rest of the time.

In order to identify the storages used (groundwater, *SWC* or *SWE*) let us consider the dynamics of the first two storages. The maximum snow water equivalent closely related to the formation of the maximum water storages did not change in total over the period from 2002 to 2015 (Fig. 2).

On average for the field routes (Fig. 2a), the growth of *SWE* was only 0.04 mm/year. A decrease in *SWE* within 1-5 mm/year is noted for the north of EPR. At the same time, there is a growth, within 1-5 mm/year as well, in the south, especially in the basin of the Volga River. For most of the forest routes (70%), located mainly in the north of EPR, a decrease in *SWE* on average by 3 mm/year is typical. For some of those forest routes where there is a growth in *SWE* (mainly in the south of EPR), it averages 1.3 mm/year. Thus, a decrease in the maximum *TWSA* in the south of EPR is not related with a reduction in *SWE*. A slight decrease in TWS in the north-east of the EPR (<5 mm/year) may be caused by a decrease in *SWE*. The near-zero *SWE* change in the north-west of EPR could not cause a growth in *TWS* (>5 mm/year) in this area.

The pattern of a change in the maximum and minimum water storages in the soil (Fig. 3) is close to that for *TWSA* - minor changes in the north and a decrease in the south.

The reduction in the minimum annual water storages in the soil (SWC_{min}), in the areas of the maximum decline in $TWSA_{min}$, does not exceed 5 mm/year. Thus, the negative trend of $TWSA_{min}$ for the Don basin of no more than 30% is due to a decrease in SWC_{min}. The remainder is probably related with a reduction in the storage of the surface (ponds and reservoirs) and ground waters. The rate of a decline in the maximum annual water storages (SWC_{max}) in the soil is already significantly higher, and it accounts for more than 50-70% of the negative trend of TWSAmax for the Don and Lower Volga River and the Caucasus region. The most

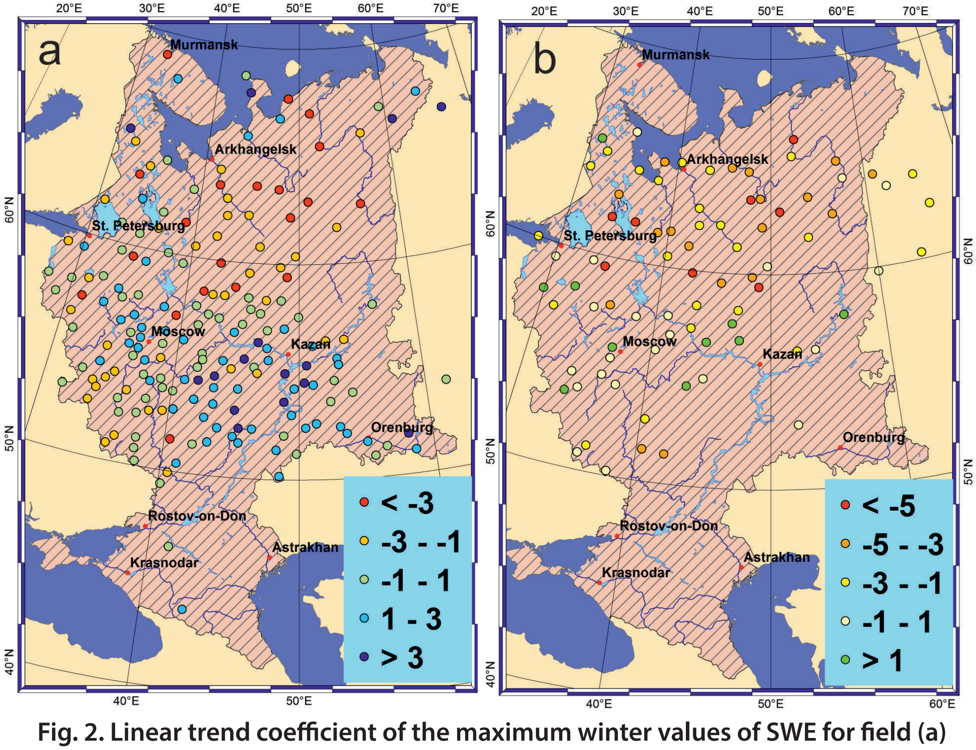

Fig. 2. Linear trend coefficient of the maximum winter values of SWE for field (a) and forest (b) routes for 2002-2015, mm/year (according to the ground-based measurements)

significant discrepancy in the dynamics of SWCmax and TWSAmax is in the basin of the Ural and partly of the Kama River, where there is practically no change in SWCmax with a decrease in TWSAmax at a rate of about 10 mm/year. The conclusion that the decline in TWSA in the southern part of EPR had an impact not only on soil waters, but also surface and ground waters, is indirectly confirmed by the fact that the impact of the initial soil moistening conditions on its subsequent dynamics affects a time interval of not more than 4-6 months (Demchenko and Kislov 2010), which is not enough to form a long period of a decline in water storages under conditions of a relatively stationary climate. The recent studies have also shown the significant values of correlation coefficients (0.46-0.83) of *TWSA* with a groundwater level for a number of wells in EPR (Savin et al. 2016).

The comparison of the accuracy of the obtained dependences *Q=f(TWSA)* in the form of a power (2) and a linear (3) function showed close results. Thus, the power dependences showed *D*=56.2 (err_{rel}=18.1%), and the linear ones - 55.4 (err_{rel}=18.7%) on average for 31 stations for the summer low water level period (July-October). The highest values of *D* and the minimum errrel were obtained for the sections of the Don River - Khutor Belyaevsky, the Don River - Stanitsa Kazanskaya and the Oka River - Murom, where *D* exceeded 80% and errrel is less than 10%. One of the reasons for the most satisfactory approximation for these sections is a significant, in comparison with the rivers of the north of EPR, range of *TWSA* that exceeds the error of *TWSA* estimation by 5-10 times. The catchment areas positively correlate with the values D of the dependences obtained (*r*=0.18 for power and 0.25 for linear dependences) and negatively with errrel (*r*=-0.51 for power and -0.53 for linear dependences), which is caused by a decrease in GRACE errors with an increase in the averaging area. The only catchment area with a shift of 1 month (*TWSA* is used as a predictor for *Q*), where the approximation accuracy has increased, is the catchment area of the Neva River, which may be related with a longer basin lag.

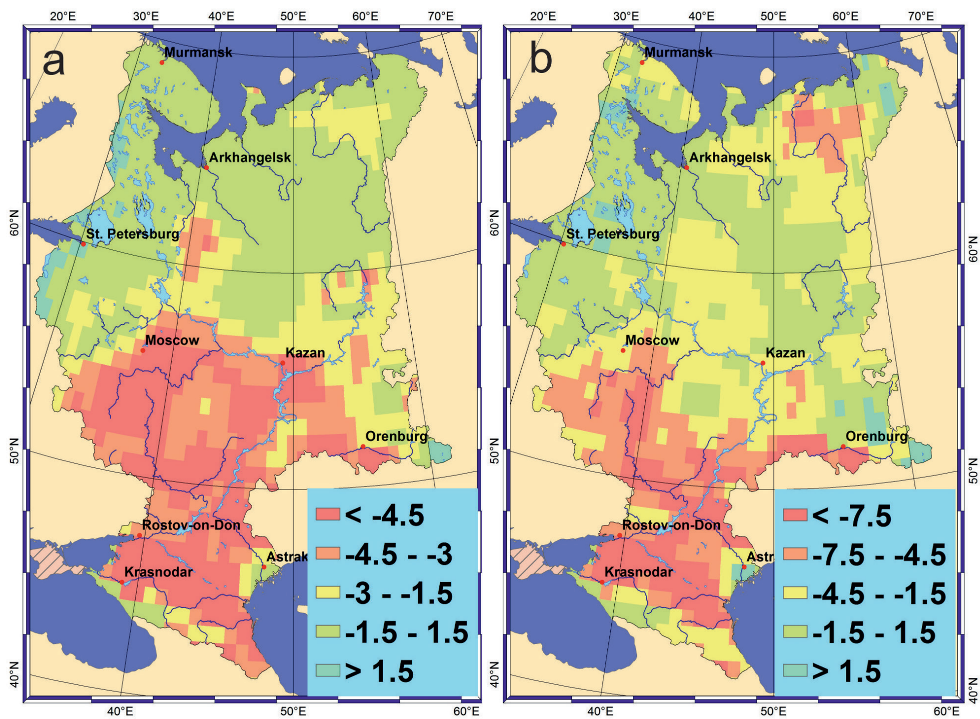

Fig. 3. Linear trend coefficient of the minimum (SWCmin) (a) and maximum (SWCmax) (b) soil water contents for 2002-2015, mm/year (according to the ERA-Interim reanalysis data)

The parameter alin in the formula (3) determines how much the runoff value increases with an increase in *TWSA* by 1 mm. The higher alin, the lower the water-retaining capacity of a catchment area, the lower the runoff autocorrelation coefficient and the lower its inertia (Frolov 2014). Therefore, the variability of terrestrial water storages (σ_{TWS}) will be lower for the catchment areas with high alin values. Thus, when using a logarithmic dependence, the determination coefficient between σ_{TWS} and alin was more than 83% (Fig. 4).

Such a dependence (Fig. 4) makes it possible to calculate a river runoff value (relative to its average long-term value) only in terms of water storages, the alin parameter can be obtained from a dependence on σ_{TWS}. However, the value of σ_{TWS} contains both the natural variability of *TWS* and the error of its estimation, which prevents the construction of a reliable dependence of σ_{TWS} on alin. Also alin is a constant value only in a certain range of *TWS* oscillations.

CONCLUSION

A number of papers were written during the functioning of the GRACE system (April 2002 - August 2017) that demonstrated the usefulness of data on terrestrial water storages for solving various problems of hydrology and other related areas. With the development of methods for processing satellite gravimetry data, as well as an increase in their accuracy due to the emergence of new models of atmosphere and ocean circulation (Dobslaw et al. 2017) and the launch of new satellites in 2018 (GRACE Follow-On), their scope of application in hydrology will extend.

The analysis of a change in *TWSA* in EPR, beginning with the second half of the 20th century, according to literature data and the GRACE data, has shown that the growth of *TWS* in EPR in the second half of the 20th century was replaced by a decline for the southern half of EPR in the 21st century, which was the most significant in the Don basin, where the rate of a decline is 14 mm/year for 2002-2015. There were no significant changes for the northern half of EPR. It has been revealed that there is a close relation between the value of terrestrial water storages and the water flow rate in the period of a summer low water level, which can be approximated by a linear dependence. The obtained dependences can be used to calculate a river runoff and its probability distribution function.

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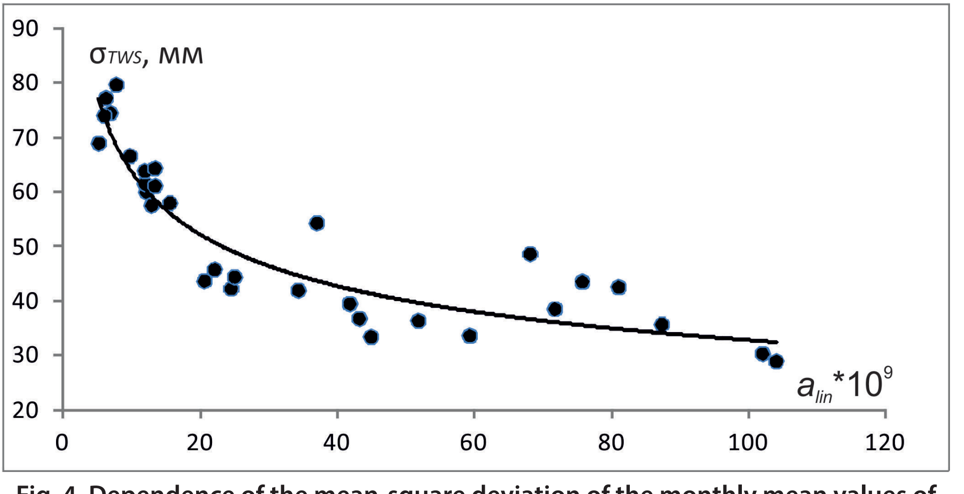

Fig. 4. Dependence of the mean-square deviation of the monthly mean values of TWSA for July-October on the proportionality coefficient between TWSA and a river runoff for 31 catchment areas

14-37-00038-P - statistical analysis), RFBR (projects No. 16-55-52008 MNT_a - flow calculations, No. 16-05-00753 A - GRACE data processing, No. 17-05-41030 RGO_a - cartographic processing).

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