Soil water regime largely determines the state of various ecosystems. Lack of moisture inevitably leads to their degradation. At the same time, excess of moisture can also have adverse effects on natural communities. For example, excess of soil moisture results in leaching of chernozems and decreases their fertility. Increased soil water content in the regions of permafrost degradation may cause bogging of the territory.

That is why it is necessary to preserve and develop the networks for observing main land water balance elements, such as soil moisture, actual evapotranspiration and pan evaporation.

**AVAILABLE OBSERVATIONAL DATA**

**Soil moisture**

Soil moisture observation network began in the USSR at the end of the 1950s. Until the middle 1980s, there were around 2000 stations operating over the Russian territory. Most of them monitored soil moisture conditions at natural grass fields (most often grasslands or wild lands) and agricultural lands with winter and spring cereal crops. Three essential criteria were observed in selecting an observation site (Guide to Hydrometeorological Stations and Posts, 1973): (i) flat surface, more than 0.1 ha; (ii) landscape and soil type are representative of the surrounding area and do not differ greatly from the prevailing landscape and soil type of the climatic zone; (iii) mean groundwater depth and its seasonal variations are typical of a large area.

According to the Guide (Guide to Hydrometeorological Stations and Posts, 1973), in the former Soviet Union (FSU) soil moisture measurements were taken once in a month during winter and every 10 days during the warm season. Also, the so-called agrophysical soil constants were calculated every five years and in case when the station was moved to another location. These constants characterize soil properties at a site. The unique thermostat-weight technique, also called gravimetric method (Guide to Hydrometeorological Stations and Posts, 1973), whereby 10-cm samples are weighed, oven-dried and weighed again, makes it possible to measure soil moisture with great accuracy, although it is rather effort-consuming. The method gives not only the total, but also plant-available soil moisture, that is portion of water that can be used by plants and involved in evaporation (total soil moisture is the wilting level).

Soil moisture is measured at agrometeorological, heat balance and water balance stations. A network of agrometeorological stations started in the 1940s-1950s and consisted of more than 2000 stations conducting specialized observations including measurements at different depths. These stations covered the whole crop growing zone of the Soviet Union. The area of risk farming (north of 60ºN) had a sufficient number of stations while most stations were located in the European Part of Russia. In the middle 1990s, not only the network but the amount of collected agrometeorological data dramatically reduced.

A network of heat balance stations was organized for collecting information on major components of the underlying surface heat balance: air temperature and humidity, wind velocity, soil temperature and moisture. This network was established in the 1960s, and by the middle 1980s there were 78 specialized stations operating in the Soviet Union. Due to the collapse of the Soviet Union and lack of finance, the network has reduced. Currently, only 39 stations located in different natural zones are operating in the territory of Russia (Fig.1).
Water balance stations were developed for studying water balance components of the land. Observations of runoff, evaporation from various surfaces, precipitation, groundwater table, soil temperature and moisture and major meteorological characteristics are performed in experimental catchments. The network began in the 1950s and in the 1980s it had 18 stations (Zavodchikov, Zhuravin, 1981). At each of the stations, three or four experimental catchments with different natural conditions were organized (Fig.2). In the middle 1980s, the network and consequently the amount of observations sharply decreased. Today, less than 10 stations operate in the Russian territory.

At the same time, all the specialized networks partially overlapped with each other, the same station being within the coverage of two or three specialized networks.

At the beginning of the 21st century, the number of stations in all networks of Russia has been reduced, and this especially pertains to observations at agrometeorological stations with natural grass fields. By the end of the 20th century, observations at natural grass fields with more than 40-year observation period were conducted at slightly more than 100 stations.

The situation is similar in the former Soviet Union states. In the Ukraine, observations have been continued mainly at agricultural fields with winter and spring crops (Robock et al, 2000, 2005). In Kazakhstan, the soil moisture monitoring program has almost been terminated: network of stationary stations is not in operation (reduced program is used for monitoring soil moisture). In Armenia, the soil moisture monitoring program has also been reduced. In Belarus, however, the network has not actually undergone changes since the 1980s (Loginov, Volchek, 2006).

At present, observational data from the whole Russian territory (more than 1500 agrometeorological stations) are collected at RIHMI-WDC. Soil moisture data for heat balance stations are collected and analyzed at the Voeikov Main Geophysical Observatory (since the middle 1990s until present). Data from separate water balance stations are currently collected at the SHI but most part of these are kept at Territorial Branches for Hydrometeorology and Environmental Monitoring (UGMSs) and are not available for scientific research.

Observation networks in China and Mongolia were developed with participation of Soviet specialists. The networks began to operate in the 1960s and still continue operating, although the number of stations has somewhat reduced. As in the FSU, the gravimetric method is used, with observations made at 10-cm layers with temporal resolution of 10 days during the warm season, and once a month during winter. Soil moisture was measured at both agricultural and natural grass and fields. In recent years, observations at natural grass fields have been considerably reduced. At present, datasets of 40 stations in China and 44 stations in Mongolia are available for research (see Robock et al., 2000 and Global Soil Moisture Bank, http://climate.envsci.rutgers.edu/soil_moisture/).
In India, soil moisture is also generally measured by the gravimetric method. At present, about 50 stations operate in the country. The stations are located at grass and agricultural fields. Observations are made once a week at several depths and only wilting level is observed (http://www.imdpune.gov.in/). Dataset for 1987-2006 is currently available.

In the United States, a network covering the whole territory is absent. There is one in Illinois where soil moisture is measured at 19 stations by the neutron probe method with time resolution of about two weeks during the growing season and once per month the rest of the year (Robock et al., 2000). There is also an experimental site in southwestern Iowa where soil moisture is measured by both gravimetric technique and the neutron probe on average twice a month.

The neutron probe technique is based on measuring slow neutrons returning to the probe from a source placed at required depths. This method allows estimation of the volumetric soil moisture content. Disadvantages of using neutron probes are in their requirement to be calibrated to soil types and zones over a period of time with different soil moisture, that they are also effort-consuming, and the need for precautions associated with handling radioactive material. Moreover, neutron probe estimations require calibration with gravimetric observations.

Most of the above data are collected in the Global Soil Moisture Bank (http://climate.envsci.rutgers.edu/soil_moisture/) (Fig. 3); detailed descriptions of soil moisture measurement techniques and currently available datasets can be found in Robock et al. (2000).

![Fig. 3. Global Soil Moisture Data Bank. Map of the distribution of the stations in the current collection (http://climate.envsci.rutgers.edu/soil_moisture/).](http://climate.envsci.rutgers.edu/soil_moisture/)

Information on soil moisture observation networks existing in Western and Eastern Europe, England, Canada, Japan, Africa and Australia is not available. Such networks are certain to exist, and it would be of use to add data from these networks to the Global Soil Moisture Data Bank. Unfortunately, available publications contain only information on soil moisture observations conducted for specific scientific projects and programs. Soil moisture is measured by various probes: neutron probe, TDR, Profile Probe, and FDR. The last three methods for soil moisture measurement include the tensiometer (a bulb of porous ceramic material placed inside the soil and connected to a water-filled tube used to measure soil moisture tension after allowing the
system to equilibrate), the gypsum block (small cylindrical gypsum blocks embedded with electrodes are buried at required depths in the soil and measure the electrical resistance, which is related to the water content), and time domain reflectometry or frequency domain reflectometry. As with the neutron probe, measurement data should be calibrated with gravimetric observations (see also Robock et al., 2000; WMO, 1994, 2008).

All the techniques can produce measurements with time resolution of several minutes to several days and at any depth (Errea et al., 2001; Diamond, Sills, 2001; Romshoo et al., 2000). Moreover, detailed description of all the existing measurement techniques is given in the Guide to Hydrological Practices (WMO, 1994).

Over the last decades of the 20th century, remote sensing data have been widely used for estimating soil moisture in the top soil layer (5-10 cm). This enables near real-time observations of soil moisture variations. Research in soil moisture remote sensing began in the middle 1970s shortly after the surge in satellite development. Subsequent research has occurred along many diverse paths. Quantitative measurements of soil moisture in the surface layer of soil have been most successful using passive remote sensing in the microwave region. Currently, data for more than a 30-year period are available (more detailed information can be found at http://www.ghcc.msfc.nasa.gov/landprocess). However, the use of such data requires careful coupled analysis of both satellite and in-situ observations of associated environmental conditions: ground level temperature and humidity, characteristics of the plant cover, precipitation and other variables. Moreover, a thorough coupled analysis of satellite and in-situ data is required for drawing a final conclusion on the possibility of using remote sensing data for estimating soil moisture in soil layers deeper than 10 cm.

**Actual evapotranspiration**

A specialized network for monitoring actual evapotranspiration was set up in the USSR during the 1950s and reached its maximum by the beginning of the 1980s. Most of the stations monitored evaporation from both agricultural and natural grass fields (grassland, wild land) (Fig. 4a).

The area selected for setting an observation site and extracting a soil core should be representative of the surrounding area with respect to the principal soil type and vegetation cover, and mean groundwater depth and its seasonal variations should be typical of a large area. Besides, the evaporation plot should be installed in an open site so as evaporation pans are not shaded and, at agricultural fields, the plot should be located at a distance not more than 3 km from the meteorological station (Guidelines on Making Observations of Evaporation from Soil and Snow Cover, 1963).

Evaporation measurements begin after the melting of seasonal snow cover and last until snow cover formation. Observations are made 6 times a month (at 5-day intervals). At agricultural fields with cereal crops, observations are made through vegetation period and sometimes after harvesting (Guidelines on Making Observations of Evaporation from Soil and Snow Cover, 1963). The so called weighing lysimeters are used on the specialized network for measuring evapotranspiration and evaporation under vegetation cover. In this case, a soil core is weighed twice on special scales and the difference indicates evaporation. At stations where intra-daily and daily evaporation rates are monitored, hydraulic evaporimeters and large lysimeters are additionally used (Guidelines on Making Observations of Evaporation from Soil and Snow Cover, 1963).

However, since the middle 1980s, the specialized network has been sharply declining. By the beginning of 2008, more than 120 stations were closed. Currently, less than 40 sites continue operating, most of them being located in the Asian part of Russia (Fig. 4b). The main reasons behind this reduction are the complex character of the observations, deterioration of the equipment and lack of personnel.

At present, observation data until the middle 1980s are only available, since reference books are no longer published and data are being collected at UGMSs.
In the former Soviet states, observation networks have also been dramatically reduced, except for Belarus where this reduction is minor - more than 30 stations are currently operating (Loginov, Volchek, 2006).

In other countries, isolated flux towers are used for measuring evaporation (both actual evapotranspiration and actual evaporation from soil) (Fig. 5). The network of flux towers was established in the 1990s and was originally designed for CO₂ measurements. Observations are made every 30 minutes or hourly. Flux towers are very unevenly distributed over the Earth, and not all of them are intended for measuring evaporation.

Actual evapotranspiration is estimated by measuring latent heat flux, sensible heat flux, soil heat and water flux, wind velocity and direction, air temperature and some others. It is calculated
by empirical formulae from these elements. Actual evaporation can also be calculated by subtracting plant transpiration from evapotranspiration, which is also possible with the use of observed characteristics. All data of observations from these two networks are accumulated at

Actual evapotranspiration (or evaporation in the case of bare soil) is measured directly by lysimeters. Lysimeters are generally installed in test fields which may include several observation sites, some of them having observation period of more than 40 years. There are such test fields in Canada, Italy, Spain, Germany, England, the USA, China and Serbia (Denich and Bradford, 2010; López-Úrrea et al., 2006, 2009; Xi Chen et al., 2008 and others.). Observations are taken with different temporal resolutions from 24 hours to 10 days. Moreover, lysimeters are used in measuring actual evapotranspiration (or evaporation) for experimental research. In this case the observation duration is not long: from one season to several years. Information on such research is available for Australia, India, Sweden and the USA (http://www.ascelibrary.org/).

There are two main types of lysimeters in use: the drainage and the weighing types. In the first case, potential evaporation is obtained as the difference between added and drained water quantity. In the second case, changes in the total weight of the soil sample are measured, whereby the real evapotranspiration during as short a time as ten minutes can be estimated. In the mechanical weighing lysimeter the soil sample is placed directly on the balance. The sensitivity will be high, if friction can be reduced using an advanced support construction. In the hydraulic weighing lysimeter the soil sample is placed in a tank floating on a fluid. Hydraulic weighing lysimeters are generally used at research fields, while at test sites both hydraulic and mechanical weighing lysimeters are used (WMO, 2008). There are a few general requirements for the location of evaporation plots: (a) the site selected for the plot should be typical of the surrounding area with respect to irrigation, soil characteristics, slope, and vegetative cover; (b) the evaporation plot should be located beyond the zone of influence of individual buildings and trees. It should be situated at a distance not less than 100 to 150 metres from the boundaries of the field and not more than three to four kilometres from the meteorological station. Soil and vegetative cover of the monoliths for inclusion in lysimeters should correspond to those of the plot (WMO, 1994).

Unfortunately, there is no any global centre to collect and store such data. Therefore, it would be expedient to establish in Russia (at the State Hydrological Institute which provides guidance for soil evaporation network) a centre for collection and analysis of data from the specialized network of Roshydromet. This would make data of direct observations available for scientific research.

**Pan evaporation**

A network of stations monitoring pan evaporation began to develop in the USSR in the 1950s and grew most efficient in the 1980s, when more than 600 stations were in operation in the territory of the USSR. Moreover, pan evaporation and soil moisture evaporation were monitored simultaneously at many of the stations, which made it possible to trace correlation between them.

Stations are divided into floating, coastal and continental depending on the size of the study basin. This enables estimating evaporation from water bodies of various sizes. The following criteria should be observed in selecting a location for observation sites: (i) evaporation plot should be flat and open to ensure that the evaporimeter is not shaded; (ii) ground water depth should not exceed 2 m; (iii) the soil composition and freeze-thaw regime should not differ from those at the meteorological station located at a distance of not more than 200 m (Guide to Hydrometeorological Stations and Posts, 1985).

Observations start from the moment of melting of snow cover, finish with the first ice cover formation and are made twice a day. Evaporation rate is measured by changes in water level within the observation interval, and then the value is summed up for 24 hours. The GGI-3000 pan developed at the SHI is used for the observations. It consists of a tank and a rain gauge.
Moreover, 20 m² and 100 m² tanks designed for studying evaporation process and its relationship with meteorological factors were installed at some stations. In the 1980s, this 20 m² tank was accepted a standard evaporimeter by WMO.

The network has decreased after the 1990s, although not so dramatically as that of soil moisture evaporation stations. About 200 stations are currently in operation in the territory of the Russian Federation, some of them having observation period of around 60 years (Fig. 6a,b).

Fig. 6a. Stations closed by 2000  Fig. 6b. Red dots – operating stations  Grey dots – stations closed by 2008

A centre for data collection and analysis has been established at the SHI. It accumulates data from all over the country. In the former Soviet states, observation networks have also reduced, except for Belarus, where it has even been slightly expanded (Loginov, Volchek, 2006).

Unlike the two previous networks, network for monitoring pan evaporation is quite dense in many countries.

In the countries of Eastern Europe, the GGI-3000 pans were probably in use until the 1990s. Unfortunately, information on the current state of these networks is not available. In the rest of the countries, Class A Evaporation Pan is mostly used for measuring pan evaporation. The evaporation pan is installed on the open-frame wooden platform (3-5 cm height), which is set and leveled on the ground in a grassy location, away from bushes, trees and other obstacles which obstruct a natural air flow around the pan, thus representing open water in an open area. Such single obstructions, when small, should not be closer than 5 times their height above the pan; for clustered obstructions, this becomes 10 times. The ground cover at the evaporation station should be maintained as similar as possible to the natural cover common to the area (WMO, 2008).

Evaporation is measured daily as the depth of water (in inches or mm) evaporates from the pan. The measurement day begins with the pan filled to exactly two inches (5 cm) from the pan top (reference level). At the end of 24 hours, the amount of water to refill the pan to exactly two inches from its top is measured. The rate of evaporation from a pan is measured by the change in level of free water surface. Often the evaporation pans are automated with water level sensors and a small weather station is located nearby (see http://en.wikipedia.org/wiki/Pan_evaporation; WMO, 2008).

The Class A Evaporation Pan is used in many countries on all continents. The USA began systematically installing Class A Pans on the network at the end of the 1950s (Lawrimore, Peterson, 2000). In Canada, the Class A Pan was also introduced at the end of the 1950s (Mukammaj, Bruce, 1960). Since the beginning of the 1960s, the Class A Pan has been in use in many countries of Europe, South America, Japan, Israel and some countries of Africa. In Thailand, this type of pan appears to be introduced later (in the 1970s) (Möller, Stanhill, 2007; Asanuma, Kamimera, 2004; Tebakari et al., 2005; and others). In Turkey and Iran, Class A Pans
have also been used for potential (pan) evaporation measurements, but since what year is uncertain (Khoob et al., 2007, Keskin, 2004).

In some countries, other types of pans are also utilized along with Class A Pans. In China, where the network has been operating since the 1960s, micro-pans are widely used as well (McVicar et al. 2007). In the British Isles, where the observation network is the oldest (since the end of the 1880s), in England and Scotland until the end of the 1960s observations were performed by the British Meteorological Office (MO) using sunken evaporation tank (Stanhill, Moller, 2008). In India (Chattopadhyay, Hulme, 1997), Australia (Rayner, 2005) and New Zealand (Finkelstein, 1973), the sunken evaporation tank was also used since the 1940s (or 1950s) until the end of the 1960s. Australian pans are fitted with bird guards that reduce the daily pan evaporation rate by about 7% (van Dijk, 1985).

The micro-pans are 0.2 m in diameter, 0.1 m deep, contain water to a depth of about 0.02 to 0.03 m, and are located on a wooden platform about 0.7 m above the ground. In contrast to the Class A Pan, the Sunken (Colorado) pan (or tank) is square, 1 m on a side and 0.5 m deep and buried in the ground to within about 5 cm of its rim. As with the Class A Pans, observations are made once in 24 hours, but for sunken pans, evaporation rate is determined by the amount of water necessary to refill it to reference level. Evaporation from a sunken (Colorado) pan can be compared with a Class A pan using conversion constants. The pan coefficient, on an annual basis, is about 0.8 (see http://en.wikipedia.org/wiki/Pan_evaporation).

There is no global centre for collection of observational data so far. Therefore, it appears very important to establish an international database (probably under the auspices of WMO) to accumulate such data. This would enable scientific community to apply data on this essential water balance element in large-scale research.

METHODS FOR OBSERVATIONAL DATA ANALYSIS

Soil Moisture

Due to diversity of soil cover and variety of its hydrophysical properties, the use of 10-day, monthly or seasonal soil moisture data in scientific research, especially in making generalizations over large areas, is a rather complicated task. The problem of using in-situ data lies also in the lack of these data for such type of analysis.

That is probably why most studies of in-situ soil moisture data are done for limited areas (river basins, regions of countries, etc.) where results do not have to be spatially generalized (Fisser, 1968; Hall, Jones, 1983; Keppeler et al., 1994; Paz Errea et al., 2001; Diamond, Sills, 2001; Kravtsov, 2008; Nazarkina, 2008). Generally, such studies use a geographic approach whereby the obtained estimates are related to certain geographic areas and regions.

Some studies deal with the analysis of particular soil characteristics (hydraulic conductivity, texture, soil horizon, heat and moisture fluxes) over relatively small areas (Herbert, 1968; Clapp, Hornberger, 1978; Boix-Fayos et al., 1998; Gusev, 2008; Kolomyts, Surova, 2010, etc.). Most of these are generally conducted under laboratory conditions.

Quite a number of studies are concerned with mathematical analysis of the structure of spatial and temporal soil moisture fields (Vinnikov et al., 1996; Entin et al., 2000; Entekhabi, Rodriguez-Iturbe, 1994; Liu et al., 2001). These studies make it possible to recognize temporal regularities of soil moisture variability as well as to evaluate areas covered by observation network. They use various statistical methods, the main being the correlation analysis.

In recent decades, soil moisture estimates obtained from aircraft and remote sensing data have been widely used for making global generalizations. A large number of recent studies deal with reliability of remote sensing data for soil moisture regime simulation (Hallikainen et al., 1985; Blyth, 1995; Jackson, Vine, 1996; Lakshmi et al., 1997; Romshoo et al., 2000; Thapliyal et al., 2003; Narayan et. al., 2006; Ramakrishnan et. al., 2006; Chandrasekar et al., 2008; Kravtsova, 2005; Pogorelov, Kiselev, 2009, etc.). These studies use various mathematical and
physical models to establish the relationship between soil moisture, underlying surface temperature, evaporation, vegetation index and albedo. In this case, correlation and regression analysis, enabling one to evaluate relationships in the soil-plant-atmosphere system, is extensively used (Blyth, 1997; Scipal, Wagner, 2004; Aires et al., 2005). Satellite data are generally validated using in-situ data gridded corresponding to a resolution of the satellite observations. Unfortunately, modern remote sensing has a number of limitations: 1) soil moisture observations are possible to perform within a rather restricted range of wavelengths and satellite inclinations and 2) soil moisture measurements are restricted in territories with closed canopy and in deserts (Choudhury, Golus, 1988; Vinnikov et al., 1999; Baup et al., 2010). In spite of the aforementioned shortcomings, satellite imagery can prove to be helpful for analysis of soil moisture patterns on both regional and global scales (Wagner, Scipal, 2000; Reichle et al., 2004; Dirnmeyer et al., 2004; Prigent et. al., 2005; Pogorelov, Kiselev, 2009).

Modern climate models, both regional and global ones, provide another possibility to investigate the global distribution of soil moisture. These are mainly water-balance models in which soil moisture is calculated from a relationship between major land water balance components, such as precipitation, evapotranspiration and runoff (Mintz, Serafini, 1984; Delworth, Manabe, 1988; Henderson-Sellers et al., 1995; Wetherald, Manabe, 1999; Yamaguchi, Shinoda, 2002; Schulte et al., 2005); in some cases also between soil and air temperature and soil infiltration (Serafini, Sud, 1987; Mintz, Serafini, 1992; Mintz, Walker, 1993; Volodyn, Lykosov, 1998, etc.). Some studies use mathematical (regression) models considering statistical dependence of moisture on major meteorological variables such as precipitation, evaporation, air temperature, wind velocity and radiation (Isham et al., 2005; Shang et al., 2007, etc). Using in-situ data of meteorological parameters for model calculations makes it possible to evaluate soil moisture distribution over large areas. To validate model estimates and parametrizations, in-situ soil moisture observations have been extensively used (Hall and Jones, 1983, Mahfouf, 1991; Robock et al., 1995-1998; Entin et al., 1999; Reichle et al., 2004; Schaake et al., 2004; Zhu et al., 2009, etc). All such studies are mostly based on correlation analysis of model estimates and observed values of soil moisture. In the process, a geographical approach is extensively applied whereby model estimates and field observation data are gridded for comparison. Although model estimations enable assessment of soil moisture on a global scale, they still have a number of substantial shortcomings: 1) model estimates largely depend on parametrizations used, 2) there exists an uncertainty in determining the atmospheric forcings, 3) observational data are lacking for validation of model estimates on a global scale (particularly soil moisture and soil surface layer temperature) (Robock et al., 2003; Prigent et. al., 2005; Guo et al., 2006). In spite of every effort to improve parametrizations used for land surface processes (Global and Planetary Change, Special Issues, 1996, 1998, 2003; Henderson-Sellers et al., 1993, 1995, 1996; Pitman, Henderson-Sellers, 1998; Mitchell et al., 2004), “…soil moisture remains one of the most difficult climatological parameters to model and . . . any computed climatology must be considered with caution.” (Douville et al., 1999).

In recent years, reanalysis data have been increasingly used for analysis of soil moisture variability on regional and global scales (Gibson et al. 1997; The ERA-40 Project, 2000; Mahfouf et al., 2000; Kistler et al., 2001; Kanamitsu et al., 2002; Uppala et al. 2005). As with climate models, reanalysis enables estimation of gridded soil moisture data while reanalysis data differ from the same shortcomings as model estimates do. Soil moisture data from ground-based measurements, satellite observations and model estimates are used (often in combination) for validation of the reanalysis data (Srinivasan et al. 2000; Douville et al., 2000; Betts et al., 2003b; Li et al. 2005; Robock et al., 2005; Aires et al., 2005; Ferranti, Viterbo, 2006; Drusch, Viterbo, 2007; van den Hurk et al., 2008). Research carried out demonstrates that reanalysis data are often inconsistent with in-situ data, but do agree with model estimates because model estimates and reanalysis data are based on different models. Unfortunately, it is premature to speak about the reliability of reanalysis data (Betts et al., 2003a; Dirnmeyer et al., 2004; Schaake et al., 2004;
IPCC, 2007) because parametrizations used in modeling land surface processes require improvement.

Major difficulty in the use of in-situ data for validation of global estimates and making generalizations lies in the scarcity of these data as well as in peculiarities and differences of datasets found in different countries. Moreover, it is difficult to take into account the diversity of soils when hydrophysical properties of certain soil types differ substantially. This is exactly why it is necessary to develop methods for in-situ data generalization.

In Europe and the United States, a methodological approach suggested in Vinnikov, Eserkepova (1989), Vinnikov, Yeserkepova (1991) is quite extensively used for analysis of in-situ data generalized over large areas. In order to avoid the problem of soil diversity in the process of spatial averaging, the authors suggested analyzing relative soil moisture (values of plant available soil moisture normalized to field capacity). Plant available soil moisture refers to the amount of water that can be extracted by vegetation cover and evaporated. This method was developed with the use of observations from the former Soviet Union and enabled one to reveal spatial and temporal soil moisture variations over the European part of the country. It was then further developed (Robock et al., 2005; Entin et al., 2000, etc.) and extended to other countries. However, this approach is also not flawless. In spite of the introduction of the relative soil moisture, the diversity of soil types still cannot be fully taken into account. Furthermore, agrophysical constants used in this approach (wilting level, field capacity and total water-holding capacity) are not always possible to obtain.

Another approach to analyzing soil moisture distribution over large areas was suggested in (Meshcherskaya et al., 1982). This approach is basically statistical. It was developed also for analysis of data from the European part of the Soviet Union. Mean monthly modular coefficients of moisture from each observation station (normalized plant available soil moisture) were averaged over the administrative districts for each month. Soil types and texture were taken into account by additional subdividing administrative districts into physiographic zones corresponding to a soil type:

- north, gray forest soil
- transition zones between gray forest soils and chernozem
- chernozem
- transition zones between chernozem and chestnut soils.

This made it possible to study statistical structure of soil moisture fields and a correlation relationship between soil moisture and various meteorological variables for all the identified zones. However, this method has not found wide application either, although the approach to analyze derived rather than observed soil moisture values was later applied, e.g. in Vinnikov, Eserkepova (1989).

The most widely adopted method for analyzing soil moisture patterns over large areas is the one suggested by L.S. Kelchevskaya (1983). This one was based on a new agrohydrological zonation. The European part of the Soviet Union was divided into nine agrohydrological zones characterized by a specific soil water regime:

- full saturation
- maximal capillary moistening
- periodic capillary moistening
- temporal overwetting
- moistening by capillary-perched and gravity moisture
- full spring wetting
- moderate spring wetting
- low spring wetting
- very low spring wetting.

This zonation takes into account not only the moistening regime of the upper 1-m (active) soil layer, but also the impact of a set of external factors which form and affect soil moisture
reserves and agrophysical soil moisture properties. The approach enabled one to compile maps of plant available soil moisture for different decades and crops. However, the method was found to be rather time and effort consuming as it took quite a long time to analyze the affecting factors. Nevertheless, some of its aspects, such as agrohydrological zonation, principles of averaging data from individual observation sites and some others, have been applied in the analysis of soil moisture variation (e.g. in research work of the All-Russia Research Institute for Agricultural Microbiology (ARRIAM)).

Along with the above methods enabling one to make assessments for the whole European Russia, there exist a number of approaches used for analysis of soil moisture patterns and forecasting soil moisture reserves for restricted areas, including individual fields. These to a certain extent use some aspects of the first two methods (see Loginov, Volchek, 2006; Speranskaya, 2009).

Therefore, direct use of in-situ soil moisture data for analysis of soil moisture patterns over large areas and on a global scale currently appears to be a rather complicated task. This is caused by a lack of observation sites and differences in observation methods used in different countries. Another aspect of the problem is concerned with absence of reliable information on agrophysical properties of soil types and, particularly, texture. That is exactly why remote sensing (mainly satellite observations) of soil moisture and associated parameters is growing in importance. Development of climate models, and especially parametrizations used for analysis and forecasting land surface processes, will in the future offer the possibility to more extensively use model estimates and reanalysis data for assessing soil moisture variations. However, in-situ data remain the most reliable source of soil moisture information so far.

**Actual evapotranspiration**

First, we need to clarify the notion of actual evapotranspiration. Both in Russia and abroad, this term refers to the sum of actual evaporation from soil surface and plant transpiration (from grass or crop vegetation cover). Evaporation from bare soil is rarely used in scientific research. There is also another concept quite commonly used in the United States and Western Europe which is the so-called reference crop evaporation or reference evapotranspiration (the reference surface is a hypothetical grass reference crop with an assumed crop height of 0.12 m, a fixed surface resistance of 70 s m$^{-1}$ and an albedo of 0.23) and under no soil moisture stress conditions. This parameter is most often used by FAO (The United Nations Food and Agriculture Organization) for calculation of irrigation and evaporation rates.

In-situ observations of soil evaporation are mostly available from the territory of Russia, former Soviet republics and some countries of Eastern Europe. However, these data being scarce, there are only a few research papers dealing with in-situ data analysis. (Kharchenko, 1959; Loginov, Volchek, 2006; Golubev et al., 2001; Golubev et al., 2002, 2003; Speranskaya, Tsytsenko, 2008).

Due to the scarcity of in-situ data, there is no consensus among Russian scientists so far on the possibilities of using such data. Some specialists suppose that applicability of soil evaporation observational data is limited by the scarcity of observation network, difficulties arising from the need to relate measured values from an individual point to large areas (e.g. river basins) and great difficulties of assessing the drainage basin active layer and its temporal variations. Thus, some pessimistic conclusions are often made that data of observations from soil evaporimeters can only be applied to small areas with a relatively homogeneous surface (see e.g. Babkin, Vuglinsky, 1982).

On the other hand, there are some opposing viewpoints. Some researchers, e.g. Krestovsky et al. (1966), suggest methods for calculating monthly evaporation totals for a watershed using data from soil evaporimeters. An opinion also exists that direct application of observations from evaporation network requires some special non-conventional methods to be developed (Koronkevich, 1990).
Lack of in-situ data results in a wide range of calculation methods applied for analysis of soil evaporation patterns. Along with simple water balance estimations based on the difference between observed values of precipitation and runoff (Vodnye Resursy, 1967; Koronkevich, 1990; Kuzmin, 1953; Hobbins et al., 2004; Walter et al., 2004), there are a large number of more complicated techniques which take into account the relationship between meteorological and physical processes affecting the evaporation.

Basic tools for assessing soil evaporation are provided by the models based on solution of heat and water balance in various forms (Penman, 1948; Thornthwaite, 1948; Budyko, 1948, Andreyanov, 1960; Budyko, Zubenok, 1961; Konstantinov, 1963; Budagovsky, 1964; Zubenok, 1976; Brutsaert, 1982). The so-called complex method combining both approaches is most often used in these models (Penman, 1948; Budyko, 1948). Therefore, it is rather formal to divide the models into water balance and heat balance ones.

Estimation of actual and reference evapotranspiration by the heat balance method is based on the account of heat balance components (net radiation, soil heat flux, thermal and humidity air characteristics) within the atmospheric boundary layer. These methods are used for assessing evaporation over small areas as well as on regional and global scales (Liang, G., 1982; Alvenäs, Jansson, 1997; Choudhury, 1999; Startseva, 1990, 1999; Droogers, 2000; Xu et al. 2006b; Zhang et al., 2007, 2009). The same method is recommended by FAO for estimation and forecasting crop water needs (reference evapotranspiration) (Allen, et al., 1998; López-Urrea et al., 2009).

Methods based on water balance of the atmospheric boundary layer take into account the dependence of evaporation on moisture availability (including soil moisture), surface air temperature and humidity and precipitation. They provide a possibility to make both regional and global generalizations (Douglas et al., 1994; Sorman, Abdelrazzak, 1994; Frank, Inouye, 1994; Bonta, Müller, 1999; Milly, Dunne, 2001; Gao et al., 2007; Yang et al., 2009).

Methods based on complementary relationship between potential evapotranspiration and actual or reference evapotranspiration have been used quite extensively for estimating actual evapotranspiration, especially in the recent decades. Values of potential evapotranspiration may be either calculated or measured with various types of evaporation pans (Kristensen, Jensen, 1975; Hsuen-Chun, 1988; Spittlehouse, 1989; Brutsaert, Parlange, 1998; Poulovassilis et al., 2001; Szilagyi et al., 2001; Hobbins et al., 2001, 2004; Xinfà et al., 2004; Khoob et al., 2007; McVicar et al., 2007). However, estimations obtained by this approach are used mainly in regional generalizations.

Along with the aforementioned methods, a large number of techniques have been developed which use different approaches to estimating actual and reference evapotranspiration on local to global scale. Such estimations use purely mathematical (based on statistical regularities), as well as physically-based, models taking into account the dependence of evaporation on various environmental factors (Kharchenko, Kharchenko, 1965; Krestovsky et al., 1966; Eagleman, 1971; McNaughton, Spriggs, 1986; Kondo et al., 1990; Kuchment, Startseva, 1991; Chen et al., 2008; Shlychkov, 2004). Techniques have also been developed for estimating actual evapotranspiration using GIS-technologies (Parajka, Szolgay, 1998).

Comparison between different models for estimating regional actual evapotranspiration and parametrizations used in the models cannot reveal whether any of the techniques are more realistic than others because no absolute validation criterion has been found so far (Konstantinov, Kozlov, 1955; Struzer, Rusin, 1956; Mihailović et al., 1995; He, Kobayashi, 1998; Sentelhas, Folegatti, 2003; Xu, Chen, 2005; López-Urrea et al., 2006; Xu, Yang, 2010; Trajković, Gocić, 2010).

However, all the aforementioned methods provide a possibility to estimate seasonal, monthly or 10-day actual evapotranspiration; estimations of daily actual evapotranspiration are made very rarely.

In recent decades, it has become possible to estimate soil evaporation on a daily (or even hourly) basis. Progress in satellite, aircraft and flux tower measurements opens up the possibility of estimating meteorological parameters, heat and moisture fluxes as well as environmental
factors used in different models with very high (sometimes up to several minutes) time resolution. Using these data in models provides an opportunity to estimate soil evaporation for various time spans and areas (Rambal et al., 1985; Taconet et al., 1986; Carlson, Buffum, 1989; Kuchment et al., 1989; Sucksdorff, Otle, 1990; Bastiaanssen et al., 1997; Granger, 2000; Ahmad et al., 2005; Loukas et al., 2005; Afanasyev et al., 2006; Muzylev et al., 2006, 2008; Wang et al., 2007; Senay et al., 2008; Tsouni et al., 2008; Guerschman et al., 2009). However, even reliability of flux tower data cannot be evaluated by a single validation criterion. Some authors compare model estimates with in-situ data, some with other model estimates or reanalysis data. It can be understood that such validation cannot provide a final conclusion on reliability of the results obtained. Furthermore, reanalysis data cannot be used as such a criterion as they itself need to be validated (Betts et al., 2003a,б; IPCC, 2007).

Therefore, limited in-situ soil moisture data makes it impossible to draw any conclusions on variability of this parameter on regional and global scales. That is why models, especially regional ones, which take into account soil and plant characteristics in different parts of the Earth, are of importance for estimating reference and actual evapotranspiration.

However, modern models are far from being perfect and need further improvement and development, which will enable to improve reanalysis data as well. At the same time, remote sensing methods also require improvement. Enhanced satellite technologies will enable more precise measurements of meteorological and environmental parameters forcing actual evapotranspiration. Nevertheless, establishing a global database of in-situ evaporation data appears to be the most urgent need because it is these data that can be used as a validation criterion for model estimates.

Pan evaporation

Pan evaporation (results of measurements by various types of evaporimeters) is often used as a characteristic of potential evapotranspiration. At the same time, potential evapotranspiration is calculated from various models based on water and heat balance of the land (Penman, 1948; Thornthwaite, 1948; Budyko, 1948, Budyko, Zubenok, 1961; Konstantinov, 1963; Zubenok, 1976; Brutsaert, 1982). Both of these types of evaporation are known to be used to define land water content.

In contrast to soil moisture and actual evapotranspiration networks, pan evaporation network is developed throughout the world. That is why the first studies into variation of evaporation from free water surface were initiated as early as at the end of the 1940s.

However, early investigations were mainly focused on methodological aspects of estimating evaporation from various water bodies and the use of available observational data (Young, 1947a,б; Zaikov, 1949). Studies into the problem of adaptation of pan evaporation observations to calculations of evaporation from water bodies of various sizes were later continued (Easterbrook, 1969; Kuznetsov, Golubev, 1971; Mautkin, 1975; Golubev et al., 1989). Most of the results were summarized by A.R. Konstantinov (Konstantinov, 1963) and W. Brutsaert (Brutsaert, 1982).

Particular interest has been focused in the recent decades on the analysis of pan evaporation data after publication of the results of investigations into evaporation trends observed over the instrumental period (Peterson et al., 1995; Golubev et al., 2001; Golubev et al., 2002). These investigations were concerned with changes in evaporation trends in the USA and Russia. Analysis of temporal tendencies (linear trends) revealed that pan evaporation has decreased since the end of the 1980’s contrary to expected increase.

The problem of pan evaporation analysis on the basis of observational data is addressed in many of the present-day studies. Basically, these are statistical methods of assessing trends (linear and other types) and analysis of the significance of such trends for selected stations (Harmsen et al., 2004; Burn, Hesch, 2007; Stanhill, Moller, 2008).

Since studies on assessing evaporation trends use data from individual observation stations, territorial generalizations are made with the use of the geographic approach whereby the
geographic location of each station is accounted. Such combination of methods enabled one to assess variations in pan evaporation for sufficiently large areas and even for countries and continents (Kolosov, Minin, 1991; Chattopadhyay, Hulme, 1997; Quintana-Gomez, 1998; Moonen et al., 2002; Asanuma, Kaiminer, 2003, 2004; Liu et al., 2004; Roderick, Farquhar, 2004, 2005; Loginov, Volchek, 2006; Zhang et al., 2007; Roderick et al., 2009a; Jeong, Kang, 2009; Jhajharia et al., 2009). The same combination of methods is used for smaller areas, e.g., basins (Lawrimore, Peterson, 2000; Milly and Dunne, 2001; Liu, Zeng 2004; Tebakari et al., 2005; Xu et al., 2006a; Wang et al., 2007; Shen et al., 2009). It is noteworthy that studies of temporal variations in pan evaporation have led researchers to an unexpected result: evaporation rates have decreased even though global temperatures are rising. This phenomenon has been dubbed the evaporation paradox.

However, using the combination of statistical methods with a geographic approach provides also some different results: either increased rates or no significant changes in pan evaporation have been observed in arid regions (Cohen et al., 2002; da Silva, 2004; Möller, Stanhill, 2007; Oguntunde et al. 2006; Tabari, Marofi, 2010).

In recent years, several new studies have been conducted that disprove earlier research results (see e.g. Kirono, Jones 2007; Jovanovic et al. 2008). The use of statistical methods made it possible to combine observations from different types of evaporation pans into a homogeneous data set and assess the validity of the trends obtained.

Many of these studies practically do not consider landscape differences. However, those that to a certain extent take the landscape differences into account, provide more realistic assessments of variations in pan evaporation and potential evapotranspiration (Golubev, Zmeikova, 1991; Golubev et al., 1994; Golubev, Tsytserenko, 1995; Golubev et al., 2003; Speranskaya, Tsytserenko, 2008). In making such generalizations, evaporation from different types of underlying surface (on the basis of pan evaporation data) is treated separately. On the basis of observational data analysis, it was convincingly proved in (Golubev, Zmeikova, 1991) that pan evaporation can be considered as a complex index characterizing annual or seasonal balance of heat and water exchange between standard water surface and the atmosphere and that it can be used as a characteristic of potential evapotranspiration from land surface.

A considerable number of studies have focused on the relationship between pan evaporation and various environmental factors (Skidmore et al., 1969; Finkelstein, 1973; Shukla, Mintz, 1982; Liang, 1982; Granger, 1989; Brutsaert, Parlane, 1998; Milly, Dunne, 2001; Roderick, Farquhar, 2002; Hobbins et al., 2004; Hongchao et al., 2005; Rayner, 2007; Cong, Yang, 2008; Roderick et al., 2009; Tabari et al., 2010). These are generally based on correlation and regression analysis. Many studies use also simple semi-empirical and empirical models describing the relationship between evaporation and meteorological variables such as wind, atmospheric transparency, solar radiation, precipitation etc. Such approach makes it possible to observe interrelationships in different natural zones of the Earth for time intervals as short as 24 hours. The obtained assessments are often used to explain the causes of variations in pan evaporation as well as to calculate potential evapotranspiration.

Methods for assessing potential evapotranspiration began to develop when evaporimeter observations started. Since then, a large number of studies have considered the problem of assessing evaporation over different areas. These generally use standard calculation schemes (Penman, 1948; Thornthwaite, 1948; Будыко, 1948; Van Bavel, 1966). However, a set of factors forcing potential evapotranspiration is different in different studies (Tanner, Pelton, 1960; Kohler, Parmele, 1967; Al-Nakshabandi, Kijne, 1974; Parlane, Katul, 1992; Linacre, 1993; Thomas, 2000; Gao et al., 2006; Shenbin et al., 2006). Some researchers put the main emphasis on wind velocity and net radiation; the others focus on atmospheric transparency, cloudiness or air temperature and humidity. Studies have also been conducted in which potential evapotranspiration is assessed making use of some new regularities extending standard parametrizations or suggesting new dependencies of evaporation on the environmental factors (Ojo, 1969; Parmele, McGuinness, 1974; Granger, 1989; Xu, 2001; Keskin et al., 2004; Rotstayn
et al., 2006). All studies dealing with calculation of potential evapotranspiration and analysis of its variability basically use regional approach whereby variations in potential evapotranspiration are considered for various natural zones and regions. Moreover, variations in potential evapotranspiration obtained by different methods are quite often validated by ground-based observation data. However, it is worth noting that the revealed trends do not always coincide with temporal variations in the measured pan evaporation.

At the same time, there is no consensus among scientists on advantages of the existing methods for assessing potential evapotranspiration. Comparison of the results of different models does not provide grounds for identifying the most reliable assessment techniques (Smith, 1965; Bordne, McGuinness, 1973; Lian, 1982; McKenney, Rosenberg, 1993; Mintz, Walker, 1993; Singh, Xu, 1997; Mitchell et al., 2004; Xu, Chen et al., 2005; Hobbins et al., 2008). Unfortunately, assessments obtained from various models differ greatly (Mitchell et al., 2004; Hobbins et al., 2008). These differences mainly depend on parametrizations which take account of different factors affecting the rate of potential evapotranspiration.

Attempts to assess potential evapotranspiration using reanalysis data (see e.g. Trubetskova, Filimonova, 2006; Fiona, Ashish, 2010) also give discrepant results. This approach will hardly be developed in the nearest future as reanalysis data still raise doubts (Betts et al., 2003a, b; IPCC, 2007).

Pan evaporation and potential evapotranspiration are the key elements in the Earth’s hydrologic cycle since their interannual variability affects the intensity of water exchange within the atmospheric boundary layer. That is why quite a number of studies deal with the analysis of changes in the hydrologic cycle over the instrumental period (Ramanathan et al., 2001; Ohmura and Wild, 2002; Liepert et al., 2004; Wild, et al., 2004). These studies generally contain model estimates based either on complex or simple general circulation models (GCMs). However, consensus does not exist in this area either on the issue of whether the hydrologic cycle is intensifying or weakening, in other words whether the variability of potential evapotranspiration is increasing or decreasing. Research results depend also on parametrizations used for assessing potential evapotranspiration.

The number of ground observation sites for monitoring pan evaporation and meteorological variables being large, their distribution over the Earth’s surface is highly uneven. That is why satellite observations and flux tower data have been increasingly used for assessing changes in potential evapotranspiration over the recent decades (Caselles et al., 1992a, b; Bastiaanssen et al., 1996; Moran et al., 1996; de Brito Bastos et al., 2000; Courault et al., 2003; Li et al., 2009). Remote sensing data are most often used as analogue to in-situ meteorological observations. This approach enables assessment of potential evapotranspiration in regions where observation network is not very well developed. However, assessments obtained using remote sensing data are not always in close agreement with in-situ data (see e.g. Li et al., 2009).

Therefore, the sufficiently dense pan evaporation observation network makes it possible to assess variations in this element of water cycle on both regional and global scales. However, situation is not so favorable with potential evapotranspiration. Diversity of existing methods results sometimes in controversial conclusions. It is necessary to make an intercomparison of assessments obtained by different methods to identify their reliability. The same refers to the models. Further development and improvement of parametrizations will enable more reliable assessments of changes in potential evapotranspiration and in the whole hydrological cycle under the condition of climate change. The same improvement is necessary also for simulation of basic fields of meteorological parameters forcing potential evapotranspiration in various reanalysis, which will also improve the reliability of the results. Application of remote sensing data for analysis of changes in evaporation seems promising. However, satellite observations need to be improved and flux tower network developed to provide more accurate measurements of meteorological parameters and characteristics of underlying surface. It would also seem appropriate to create a global database of in-situ observations at the network of evaporation stations, which will probably require guidance from WMO. This would make global observation
data available for analysis and help us extend our knowledge of this element of the global water cycle.

NEW METHOD FOR ASSESSMENT OF EVAPOTRANSPIRATION

The problem of evaluation of actual evapotranspiration arose almost simultaneously with the analysis of the land water balance (Penman, 1948; Thornthwaite, 1948; Budyko, 1948). Penman (Penman, 1948) asserted that for insufficient water supply, actual evapotranspiration is proportional to pan evaporation (or potential evaporation), with the intensity relying on water availability. In the early 1960s (Bouchet, 1963) it was shown that between pan evaporation ($ET_p$) and actual evaporation ($ET_a$) there is a complementary relationship (Fig. 7): decrease in areal evaporation compensates exactly the increase in potential evaporation. These results initiated researches on assessment of actual evapotranspiration from pan evaporation (or potential evaporation), which are based on two different postulates: 1) actual evapotranspiration is determined in intensity by potential evaporation and changes proportionally as a function of the latter (Penman, 1948) and 2) actual evapotranspiration determines pan evaporation (or potential evaporation) and varies inversely with the latter (Bouchet, 1963). In the last mentioned case, under conditions of unlimited water availability, the values of both evaporation types become equal and the value is referred to as “the wet environment evapotranspiration” ($ET_w$) (Bouchet, 1963; Morton, 1965).

![Conceptual Complementary Relationship](image)

Figure 7. Schematic representation of the complementary relationship in regional evapotranspiration (based on Bouchet, 1963; Morton, 1965).

However, the idea suggested by Penman (Penman, 1948), was accepted only by some scientists (see e.g. Xinfa et al., 2004). Therefore majority of researches are devoted to the analysis of Bouchet's hypothesis. To develop this hypothesis, a large number of models to estimate actual evapotranspiration were created. The most popular modern models directly using the complementary relationship approach are: the advection-aridity (AA) model of Brutsaert and Stricker (Brutsaert and Stricker, 1979), the CRAE model of Morton (Morton, 1983), and the GG model proposed by Granger and Gray (Granger and Gray, 1989).

The development of such models and new methods for evaluation of all the elements that constitute the complementary relationship and affect it for various time intervals and different areas allowed to confirm the feasibility of Bouchet's hypothesis (Priestley, Taylor, 1972; Parlange, Katul, 1992; Brutsaert and Parlange, 1998; Hobbins et al., 2001a, b, c; Szilagyi, 2001; Ramirez, et al., 2005 and others). Furthermore, this hypothesis was used, for example, to explain the observed trends of evaporation related to the ongoing climate change (Brutsaert and Parlange, 1998; Roderick, Farquhar, 2002; Hobbins et al., 2004; Ohmura and Wild, 2002).
Some papers are published in which the complementary relationship is used in a form different from the classical one (Granger, 1989). In addition, in works (Doorenbos, Pruitt, 1977; Snyder, 1992; Allen et al., 1998; Raghwanshi, Wallender, 1998; Sentelhas, Folegatti, 2003; Szilagyi, 2007) various approaches and methods for assessment not of evaporation values but of the pan coefficients (K_p) between pan evaporation and actual evapotranspiration are discussed. However, in almost all studies the coefficient K_p values are treated as fixed numbers. At the same time, there are studies in which relationship between actual evapotranspiration and maximum evapotranspiration/potential evaporation is approximated by different functional dependencies on: 1) Budyko aridity index (Budyko, 1956) (Fedorov, 1965, Kuz’min et al, 1966), 2) the difference between potential evapotranspiration calculated by the complex method (Budyko, Zubenko, 1961) and observed precipitation (Golubev et al, 1961), 3) the soil moisture content (Poulavassilis et al., 2001), 4) the combined effect of several factors (Hsuen-Chun, 1988).

This very approach based on a functional relationship between actual evaporation/evapotranspiration and pan/potential evaporation was the basis for the proposed method for assessment of actual evapotranspiration with the use of pan evaporation observations.

**Description of the proposed method**

The proposed method for assessment of evapotranspiration over large basin is based on the assumption that there is the complementary relationship between pan (or potential) evaporation and actual evaporation (or actual evapotranspiration). Availability of observations on evaporation of both types defines the limitation of the approach application. However, using minimal amount of observations on environmental elements (i.e. precipitation, pan and actual evaporation) and the absence of complex theoretical constructions allow to suggest that the applied empirical relationships are the most suitable to describe existing natural feedbacks between elements of the land water cycle. Each estimation model for the water cycle elements introduces some uncertainties as a result of using different parameterizations and assumptions. On the other hand, the program of in situ observations is being reduced and, consequently, the ability to use in situ data to develop more realistic parameterizations is reduced as well. This is precisely why the use of empirical and semi-empirical relationships seems to be more justified.

Within the framework of the proposed method the relationship between pan evaporation (or potential evaporation) and actual evaporation (or actual evapotranspiration) is considered as a variable functional relationship:

$$\frac{E_a}{E_p} = f(E_p - P),$$  \hspace{1cm} (1)

where $E_a$ is actual evaporation/evapotranspiration from virgin soil (mostly meadow), $E_p$ is pan evaporation (or potential evaporation), $P$ is precipitation.

The difference between pan evaporation and the amount of precipitation during the same period ($E_p - P$), or visible evaporation (a concept long used in oceanography) allows to take into account the current moisture conditions for each given month. In the proposed method visible evaporation directly characterizes not only pan evaporation, but it is also an indirect indicator of energy resources losses during the process of evaporation (or latent heat of evaporation) (Konstantinov, 1963; Kharchenko, Kharchenko, 1965). Furthermore, visible evaporation value allows to evaluate resources of surplus moisture in the region. A positive value of visible evaporation indicates a deficit in the regional water budget, and water demand by the atmosphere exceeds precipitation (so-called 'dry' conditions are perceived). When precipitation exceeds pan
evaporation, visible evaporation is negative (which corresponds to 'humid' conditions). The more negative the visible evaporation, the wetter the study area.

The ratio of two types of evaporation (coefficient $K_1$) is well described by an exponential equation of the form $K_1 = ae^{-nx}$, where $x$ is visible evaporation, and $a$ and $n$ are empirical coefficients, so the calculation formula changes to:

$$E_a = E_p \times K_1.$$  \hspace{1cm} (2)

It should be noted that pan evaporation data used are not the direct observation data obtained by the GGI-3000 pan but the evaporation, corrected for 20 m$^2$ tank according to the formula proposed in (Golubev, Uryvaev, 1983, Golubev et al, 1989; Golubev, Tsytsenko, 1992). This allows to estimate potential evaporation from a sufficiently large area and to determine the ratio of two types of evaporation more reliably.

To estimate reference evaporation from agricultural fields additional coefficient is introduced, which characterizes the ratio of evaporation from meadow and different crops:

$$E^{*a} = E_p \times K_1 \times K_2,$$  \hspace{1cm} (3)

where $E^{*a}$ is evaporation from various crops;

$E_p$ is pan evaporation;

$K_1$ is the ratio between actual evaporation from meadow and pan evaporation;

$K_2$ is the ratio between actual evaporation from meadow and reference evaporation from crops.

Such approach to actual evapotranspiration assessment was made possible only because over the territory of Russia there is a well developed network of specific stations performing observations both on pan evaporation and actual evaporation. Furthermore at many stations of actual evaporation measurement these observations are carried out not only at plots with natural (mostly meadow) vegetation, but also at fields with different agricultural crops (i.e. reference evaporation).

The availability of such data makes it easy to estimate transition coefficients $K_1$ and $K_2$ for different soil types and environmental conditions. Thus, the first estimates of the ratio between actual evaporation from natural vegetation and various agricultural fields were obtained in Russia in the late 1960s and early 1970s (Kharchenko, 1966; Avdeev, Fedorov, 1971). Similar coefficients are also recommended by FAO for reference evaporation calculation (Allen et al., 1998).

With irrigation, reference evaporation is calculated as $E^{*a} = (E_p \times K_1)/1.75$ (Kharchenko, 1975).

Thus, basing on these relationships, it is possible to calculate reference evaporation/evapotranspiration from different types of agricultural crops for each month of the warm period using only the results of regular observations on actual and pan evaporation and precipitation.

To calculate actual evapotranspiration from a certain area (e.g. river basin) it is necessary to assess evaporation from different types of the landscapes and land-use elements.

Thus, evapotranspiration from the forest, which occupies substantial part of Russia, is estimated by the method described in (Golubev, Tsytsenko, 1995). Its essence is that evapotranspiration from the forest is correlated to pan/potential evaporation. In such case evapotranspiration is calculated as follows:

$$E_f = E_p \times [(1-A_f) / (1-A_w)],$$  \hspace{1cm} (4)
where \( E_t \) is evapotranspiration from the forest,
\( A_f \) and \( A_w \) are albedo of the forest and water respectively,
\( E_p \) is the same as in equations (1) and (2).

The value of \( A_w \) is taken equal to 5%. For coniferous and deciduous forests different values of albedo \( (A_f) \) are used in accordance with (Pivovarova, 1977; Albedo..., 1981). These values also vary in different parts of the European part of Russia (Table 1).

Table 1. Mean monthly values of the albedo (in %) for different types of forests over the European part of Russia.

<table>
<thead>
<tr>
<th>ETP Zones</th>
<th>Months</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
<th>11</th>
<th>12</th>
</tr>
</thead>
<tbody>
<tr>
<td>coniferous forest</td>
<td>North of 60N</td>
<td>7/60²</td>
<td>28</td>
<td>14-18</td>
<td>14-18</td>
<td>14/16²</td>
<td>15-17</td>
<td>31-34</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>South of 60N</td>
<td>7/60²</td>
<td>16</td>
<td>14-18</td>
<td>14-18</td>
<td>14/16²</td>
<td>16</td>
<td>23</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>all zone</td>
<td></td>
<td>46</td>
<td>45</td>
<td>45</td>
<td>35</td>
<td>14</td>
<td>14</td>
<td>14</td>
<td>14</td>
<td>15</td>
<td>15</td>
<td>45</td>
<td>45</td>
</tr>
<tr>
<td>deciduous forest</td>
<td>West of 45 E</td>
<td></td>
<td>25</td>
<td>15-19</td>
<td>16-18</td>
<td>26</td>
<td>15-18</td>
<td>18-19</td>
<td>17-19</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>East of 45 E</td>
<td></td>
<td>35</td>
<td>15-19</td>
<td>17-20</td>
<td>19</td>
<td>17-18</td>
<td>17-19</td>
<td>17-19</td>
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<tr>
<td>all zone</td>
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<td>46</td>
<td>46</td>
<td>45</td>
<td>25</td>
<td>13</td>
<td>18</td>
<td>19</td>
<td>19</td>
<td>20</td>
<td>32</td>
<td>45</td>
<td>46</td>
</tr>
</tbody>
</table>

Note: all zone – data from (Albedo…, 1981). ETR zones – data from (Pivovarova, 1977); index “1” indicates the Western part of the zone, index “2” indicates the Eastern part of the zone.

Bogs also occupy a considerable area, especially in the regions of sufficient moisture (according to Budyko, 1956). Evapotranspiration from lowland and raised bogs is estimated in accordance with (Romanov, 1961, 1962; Krestovsky, Knize, 1993): evaporation from lowland and raised bogs is taken equal to the evaporation from meadow \( (E_a) \) adjusted for latitudinal component taken from (Romanov, 1961, 1962).

Table 2. The ratio of evapotranspiration from bogs to evapotranspiration from dry valleys (meadows) for different latitudinal zones of the European part of Russia.

<table>
<thead>
<tr>
<th>Data source</th>
<th>Latitudinal zones</th>
<th>North of 64 N</th>
<th>60-64 N</th>
<th>56-60 N</th>
<th>54-56 N</th>
<th>52-54 N</th>
<th>South of 52 N</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>raised bog</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Romanov, 1961</td>
<td></td>
<td>1.22-1.60</td>
<td>1.16-1.35</td>
<td>1.00-1.16</td>
<td>0.97-1.03</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Romanov, 1962</td>
<td></td>
<td>1.10-1.31</td>
<td>1.05-1.15</td>
<td>1.00-1.12</td>
<td>0.97-1.03</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td></td>
<td>lowland bog</td>
<td></td>
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<tr>
<td>Romanov, 1961</td>
<td></td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>0.99-1.11</td>
<td>0.99-1.00</td>
<td>-</td>
</tr>
</tbody>
</table>
For months with air temperature below freezing, evapotranspiration from all types of land surface (excluding wooded areas) is determined by calculation according to the formula of P. P. Kuz’min (Kuz’min, 1953)

\[ E_a = 0.34 \sum d, \]  
and during the snowmelt according to the formula of A. N. Postnikov (Postnikov, 1977)

\[ E_a = 0.44 \sum d, \]  
where \( d \) is mean daily values of air saturation deficit for the definite month.

The use of equation (6) allows not to take into account the definite dates of snow cover loss since it is also used to calculate the evaporation from the snow surfaces, land and water as puddles.

Guidelines for the implementation of the proposed method

The preliminary stage of research

1. Before using observations on pan evaporation it is necessary to correct original data for uniform time interval. For this it is necessary to:
   A. determine length of the warm period basing on mean monthly air temperature (months with \( t^0 C > 0 \));
   B. correct original data on pan evaporation for real warm period by the method proposed in (Golubev et al, 1989):

\[ E_m = E_n \times \frac{d_m}{d_n}, \]  
where \( E_m \) is the corrected pan evaporation value for the whole warm period, 
\( E_n \) is original (measured) pan evaporation for the actual observation period, 
\( d_m \) is the sum of mean daily air saturation deficit for the whole warm period, 
\( d_n \) is the sum of mean daily air saturation deficit for the actual observation period.

In practice, the application of this equation (7) is divided into several stages:

- At the first stage the value of the total evaporation for the whole warm period is calculated. In this case \( E_m \) is the total sum of pan evaporation for the whole warm season, 
\( E_n \) is the sum of pan evaporation for the actual observation period, 
\( d_m \) is the sum of mean daily air saturation deficit for the whole warm period, and \( d_n \) meaning retains the same.
- Then the difference between \( E_m \) and \( E_n \) is distributed between the restored months taking into account the air saturation deficit for each month. In this case \( E_m \) is the corrected value of pan evaporation for the definite month, \( E_n \) is the difference between the corrected total sum of pan evaporation for the whole warm period and value of pan evaporation for the actual observation period, 
\( d_m \) is the air saturation deficit for the definite restored month, and \( d_n \) is the sum of air saturation deficit for all the restored months.

2. When monthly values of pan evaporation are obtained for all months of the warm period, it is necessary to corrected these GGI-3000 pan observations for 20 m² tank data according to the formula developed in (Kuznetsov, 1970; Golubev, Uryvaev, 1983, Golubev et al, 1989):
\[ E_p = E^*_p \times [0.725 + (2 + 24.3n - 0.6\varphi) \times 1/E^*_p - 3.6 \times 10^{-4} \times E^*_p / n], \tag{8} \]

where \( E^*_p \) is the total sum of pan evaporation from the GGI-3000 pan observations for the whole warm period (with the restored months being taken into account),

\( n \) is the number of months per year with positive mean monthly air temperature (the length of the warm period),

\( \varphi \) is latitude of the observation point with pan evaporation, that indirectly allows to take into account changes in the incoming solar radiation amount in different latitudinal zones.

Final distribution of the total (for the whole warm season) evaporation \( E_p \) between the months of the warm period is also executed by the equation (7). In this case the meaning of variables changes again: \( E_m \) is the evaporation in the definite month, \( E_n \) is equal to \( E_p \) (Eq. 8), \( d_m \) is the mean daily air saturation deficit for the analyzed month, and the \( d_n \) is the total sum of mean monthly air saturation deficit for the whole warm period.

The use of air saturation deficit in equations (7) and (8) allows to account moisture conditions within the atmospheric surface layer.

3. On the base of satellite images or published sources the area occupied by different landscapes and land-use elements and their changes from year to year are assessed for the study period. It is important to identify the wooded areas, areas of agricultural land and of open water surface (rivers, lakes, reservoirs, etc.).

4. The ratio of actual evaporation from natural (mostly meadow) vegetation and from different agricultural crops is estimated, i.e. the \( K_2 \) coefficient in equation (3) is determined. It is preferable to compare the data for 10-days time interval, but the monthly values can also be used. The availability of observations on actual evaporation determines the division of agricultural land according to different types of crops. If these observations are unavailable, values of transition coefficients obtained for various crops and soil types within the European part of Russia can be used instead (Kharchenko, 1966; Avdeev, Fedorov, 1971, Golubev et al, 2003).

Table 3. The ratio of evaporation from the crops (\( E^*_a \)) to the evaporation from natural vegetation (\( E_a \)).

<table>
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<tbody>
<tr>
<td></td>
<td></td>
<td>Derno-podzolic soil</td>
<td>Ordinary Fertile Carbonate Leached</td>
</tr>
<tr>
<td>Spring wheat</td>
<td>1.00-1.06</td>
<td>1.10</td>
<td>1.16-1.18</td>
</tr>
<tr>
<td>Winter wheat</td>
<td>1.11</td>
<td>-</td>
<td>1.05-1.08</td>
</tr>
<tr>
<td>Barley</td>
<td>-</td>
<td>-</td>
<td>1.00-1.13</td>
</tr>
<tr>
<td>Winter rye</td>
<td>0.96</td>
<td>1.01</td>
<td>0.89-0.92</td>
</tr>
<tr>
<td>Spring rye</td>
<td>1.00-1.08</td>
<td>-</td>
<td>1.13-1.15</td>
</tr>
<tr>
<td>Oat</td>
<td>-</td>
<td>0.96</td>
<td>-</td>
</tr>
<tr>
<td>Cultivated herbage</td>
<td>-</td>
<td>1.00-1.09</td>
<td>1.18-1.22</td>
</tr>
<tr>
<td>Potato</td>
<td>-</td>
<td>0.98</td>
<td>0.97</td>
</tr>
</tbody>
</table>
The main stage of the actual evapotranspiration assessment

1. Basing on corrected data on pan evaporation \( E_p \) the regions with similar temporal changes in pan evaporation are selected over the study territory. Then data on pan evaporation are averaged by the method of optimal averaging developed by R.L. Kagan (Kagan, 1979) within each selected region. As the result, for each region there is one time series of pan evaporation characterizing its temporal changes.

2. In each selected region stations with actual evaporation observations are selected for those time intervals where parallel observations on both types of evaporation are available. Then sub-regions are determined where temporal changes in actual evaporation are similar and all available data are averaged within each sub-region by the same method.

3. For each region and sub-region empirical coefficients \( a, n \) and value of transition coefficient \( K_1 \) (\( K_1 = ae^{-nx} \), where \( x \) is the visible evaporation) are estimated, and actual evaporation is assessed by equation (2) for different intervals of range of visible evaporation values. The more intervals in the range of visible evaporation values are used, the more accurately transition coefficients \( K_1 \) are determined and the more accurately actual evaporation is assessed.

Based on these relationships it is possible to assess actual evaporation for any period and within different natural zones using only data on pan evaporation, precipitation and air saturation deficit.

This method allows to assess actual evaporation for each month of the warm period.

Actual evapotranspiration assessment for river basin or any selected region

Under determining the total actual evapotranspiration for river basin or any selected region, firstly monthly actual evapotranspiration from natural herbage (virgin land, meadow) is calculated for all months of the warm season by equation (2) for analyzed years. For months of the cold season equation (5) is used, except for the last cold month of spring for which equation (6) is used.

Then, using equation (3) and transition coefficients \( K_2 \) (see Table 3), the values of reference evapotranspiration from various crops without irrigation and from crops with irrigation (as \( E^{*}_{a} = \frac{E_p \times K_1}{1.75} \)) are estimated for all months of the warm period. For months of the cold season equations (5) and (6) are used as mentioned above.

Dividing the study territory into latitudinal zones in accordance with (Romanov, 1961, 1962) and by the ratio of areas with different bogs (raised and lowland), the actual evapotranspiration from bogs is calculated basing on data on actual evapotranspiration from natural herbage. For this purpose transition coefficients from (Romanov, 1962) shown in Table 2 are also used.

Within the study territory zones of coniferous, deciduous and mixed forests are selected. Areas occupied by mixed forests are distributed between the zones of coniferous and deciduous forests in the ratio 1:1 up to 45 E and 2:1 to the west of 45 E. Then actual evapotranspiration from the wooded area is estimated by equation (4) separately for coniferous and deciduous forests. For winter and summer periods \( A_f \) values from (Albedo ..., 1981) are used, for the transition seasons (spring and fall) - from (Pivovarova, 1977) (see Table 1).

At the final stage the total evapotranspiration from the river basin or the analyzed territory is assessed. For this purpose actual evapotranspiration from different types of landscape and land-use elements is summed taking into account the part of the total area of the study basin or

| Fallow land | 1.50 | 0.86 | 0.82 | 0.82-0.91 | - | 0.81-0.85 | 0.93 |
territory occupied by each element as weighing coefficient. This approach allows not only to evaluate the total actual evapotranspiration from the analyzed area, but also to assess the contribution of evapotranspiration from different landscape and land-use elements into the formation of its total value.

The proposed method allows to estimate actual evapotranspiration based on empirical relationships with the use of minimum amount of observations. Such approach can provide the account of the most important feedbacks between the elements of the land water cycle and reliability of obtained assessments.

Estimations of the total actual evapotranspiration from the river basin are also valuable and can be used for assessment of the state of the environmental water supply and its evolution.

**A CASE STUDY FOR THE EUROPEAN TERRITORY OF RUSSIA**

A case study for two or three countries was supposed to be performed on the base of the developed method for actual evapotranspiration estimation. However, the network of stations on actual evaporation measurement exist mainly in the countries that were members of the former USSR. There is information about regular observations on actual evaporation in China; however, reliable information about these observations length is not available at present. Furthermore, the method was developed for the Volga basin, and was used for the Don basin later on. However, each basin was analyzed separately (Golubev et al, 2003; Speranskaya, Tsytsenko, 2008). It appeared to be interesting to test this method applicability for the analysis of the joint territory of two basins with the use of common parameterizations of the relationship between pan/potential evaporation and actual evaporation from natural (mostly meadow) herbage for different environmental conditions.

For this reason it was decided in 2011 to restrict case study by the territory of European Russia, with wide range of variation in environmental conditions, to test reliability of obtained estimates and to assess possibility of the proposed method use in such conditions. For this purpose part of European territory of Russia was selected where the Volga and Don river basins are located.

**Assessment of actual evapotranspiration on the base of regular pan evaporation**

The Volga basin occupies the large territory from boreal forest (taiga) up to the zone of dry steppe. The Don basin is located mainly in the zone of fertile steppe. Only the very north of the basin is located within the zone of deciduous forests, and in the south-east the zone of dry steppe is fixed. These two basins differ sharply in their natural conditions and their joint territory is a good test site for investigation of the developed method applicability.

Because of large territory of the Volga basin, it was additionally divided into three sections: Upper Volga - up to Cheboksary (Volga-1), Middle and Lower Volga - from Cheboksary to the Caspian Sea (Volga-2) and the Kama river basin (Kama). The Don basin was considered as a whole (Figure 8).

In the Volga basin and its adjacent territory there are 53 stations which monitor pan evaporation and 41 stations with actual evaporation measurement. Furthermore, 38 of them perform parallel observations on pan and actual evaporation.

On the territory of the Don basin and in the immediate vicinity (part of the adjacent territory is located in the Volga basin) 32 stations with pan evaporation observations and 21 stations measuring actual evaporation are located. For 8 of them parallel observations on pan and actual evaporation are available.

In general, over the study territory 76 stations monitor pan evaporation and 43 stations – actual evaporation. For 25 of them parallel observations on pan and actual evaporation are
available (Figure 9). At some stations observations on actual evaporation are performed not only for natural herbage but also for different agricultural crops.

Data from available stations characterize all natural zones selected over the territory of two basins.

Data on actual evaporation for both basins are mostly available from May to October. That is why it was decided to use this period to estimate transition coefficient $K_1$ ($K_1 = ae^{-nx}$, where $x$ is the visible evaporation, $a$, $n$ are empirical coefficients) between mean actual and pan evaporation for the warm season.

Primarily original data on pan evaporation from GGI-3000 pan were corrected for the selected warm period (May-September) by equation (7), and then for data of 20 m$^2$ tank with the use of equation (8). Corrected data characterize pan evaporation not from single GGI-3000 pan but from reservoir of medium size (up to few thousand km$^2$) (Golubev et al, 1961).

To determine transition coefficients $K_2$ (the ratio between actual evaporation from the natural herbage and evaporation from different crops) for the selected warm season, in equation (3) data on reference evaporation from different crops for 23 stations in both basins were used. In case observations were missing, estimates given in Table 3 were used (Kharchenko, 1966; Avdeev, Fedorov, 1971, Golubev et al, 2003).

Landscapes and land use elements in these basins and their temporal changes were estimated by the published sources; and for this reason data collection was based on the territory division into the administrative units (state regions and republics). These elements consist of agricultural land, forested areas, open water surface, bogs, and urban area. The agricultural land
include arable lands under different crops, haylands, pastures, multi-year plantations, and fallow land. In its turn, the arable land was subdivided into areas occupied by spring and winter cereals, industrial crops, potatoes, vegetables and forage crops. Forested areas were additionally subdivided into coniferous and deciduous ones according to the species composition. Open water surface included the water areas of natural lakes, artificial ponds and reservoirs. Bogs were divided into raised and lowland. Urban area was not considered in the further analysis since assessment of evaporation from these areas is in itself a difficult problem. Furthermore, such assessment requires information about numerous factors affecting the value of such evaporation. That is why the proposed method is used to estimate the total evapotranspiration for non-urbanized environment.

For each selected region (Fig. 8) areas of different landscapes and land use elements were estimated by summing their territories identified within the boundaries of the state regions and republics located in the study area. If the territory of the republic or state region was located just on the border between the two regions and its area was divided between them in some ratio (e.g. 20% and 80%), then territories of all types of landscapes and land-use elements were distributed between the neighboring regions in the same proportion. The collected materials made it possible to monitor the dynamics of area changes for all landscapes and land-use elements from the 1950s to 2008.

Analysis of temporal changes (1955-2008) in pan evaporation in the united area of two study basins allowed to select 10 regions with similar evaporation changes (Figure 10).

![Regions characterized by the specific features of the temporal changes in pan evaporation.](image)

In each selected region all available data were averaged over its territory. To reduce the
effect of irregular distribution of pan evaporation stations over the study basins, the method of optimal averaging with weights’ normalization was used (see details in (Kagan, 1979; Groisman, Lugin, 1985, Groisman et al, 1986)). Moreover, weighing coefficient for each station is proportional to distance between this and nearby stations, and the sum of all weights is supposed to be equal to the unit. This approach allows to obtain unbiased estimates of values averaged over the territory, greatly reducing the effect of irregular stations’ distribution.

For each selected region (Fig. 10) the data on actual evaporation were collected. If there was one station within the region or no stations at all, the stations located in immediate vicinity were used for the averaging. This is to some extent reduced the accuracy of suggested approximations, but allowed to characterize all selected regions. Then these data were processed just as data on pan evaporation with selection of sub-regions with similar changes and future averaging.

Based on pan and actual evaporation data for each region the approximations for the estimation of transition coefficient \( K_1 = a e^{-nx} \) (\( x \) is visible evaporation, and \( n, a \) – empirical coefficients) were obtained for different intervals of the visible evaporation values range. For this purpose the whole range of possible values of visible evaporation was divided into parts for more accurate account of specific features of relationship between pan and actual evaporation for its positive and negative values.

At assessment of visible evaporation data on monthly precipitation were used that are available from the archive created by RIHMI-WDC. These data are available on its Website [http://meteo.ru/climate/sp_clim.php](http://meteo.ru/climate/sp_clim.php) and on Website of US NOAA National Climatic Data Center [http://lwf.ncdc.noaa.gov/oa/climate/climatedata.html](http://lwf.ncdc.noaa.gov/oa/climate/climatedata.html). Data on air saturation deficit were collected from the regular meteorological Reference-books.

Using transition coefficients \( K_1 \) for each selected region value of actual evaporation from natural (mostly meadow) herbage was estimated by equation (2). Time series of mean for the warm period values of observed and calculated actual evaporation for 1964 to 1990 are given in figure 11. As actual evaporation observations are unavailable from 1986 due to their publication end the restored time series are limited by 1990.

Obtained results allow to assert that the use of the routine observations on pan evaporation is possible for estimation of actual evapotranspiration. However, the original data require some preliminary processing. It should also be mentioned that the more intervals in the range of visible evaporation values are used, the more accurate are estimates of transition coefficients \( K_1 \) are obtained, and, consequently, assessments of actual evapotranspiration as well.

In contrast to previous studies (Golubev et al, 2003; Speranskaya, Tsytsevskiy, 2008), where for Volga and Don basins their own approximations were developed to estimate the transition coefficient \( K_1 = a e^{-nx} \), in the present Case study more universal approach is used. Parameterizations are developed not for the definite basin and soil areas, but for the regions with the similar changes in pan and actual evaporation. Such parameterization developed for the large area comparable, for example, to the European territory of Russia, can be further used for assessment of actual evapotranspiration in the wide range of scientific researches.
Assessment of the total actual evapotranspiration from the Volga and Don rivers’ basins

Assessment of the total actual evapotranspiration from the study basins consisted of several stages and included two different blocks: the first for agricultural land, and the second for the rest of landscapes and land-use elements.

Agricultural land. The total actual evapotranspiration was estimated for the selected regions (Fig. 10). At the first stage parts of the study basins (in %) belonging to each selected region were determined (Fig. 8). Then the area of agricultural land was divided between the selected regions according to obtained percentage. In each region, the area of agricultural land was additionally divided into parts with different elements (arable lands under different crops, haylands, pastures, multi-year plantations, fallow land) and some weighing coefficient was assigned for each part. Moreover it was assumed that the sum of assigned weights was equal to the unit. For crops on arable land the sum of weighing coefficients was also equal to the unit.

Using equation (3) and the coefficients values $K_2$ (both new calculated and listed in Table 3) actual evaporation from various elements of agricultural land was estimated. At this, the prevailing type of soil at the observation plots and in each selected region was taken into account. The value of transition coefficients $K_2$ in Table 3 was taken as the mean from given ranges of their changes.

At first the weighed average (considering the defined weighing coefficients for different agricultural lands) reference evaporation from each element of agricultural land was calculated within the selected region (Fig. 10). Then the total actual evapotranspiration from agricultural land was estimated for the study basins, taking into account the specific contribution of all the regions located within them.

Parts of the study basins, occupied by forests, bogs and open water surface. The division of the study basins into parts corresponding to the selected regions (Fig. 10) was not
done. The total actual evapotranspiration was estimated for the whole river basin or sub-basin, taking into account the weighing coefficients for each landscape and land-use element which values were assumed to be proportional to their areas).

To estimate the evaporation from bogs each of the study basins was divided into zones, defined in (Romanov, 1961, 1962), corresponding to different ratios of raised and lowland bogs’ areas. For each of these zones actual evapotranspiration for the warm period was estimated, taking into account defined weights of each bog type, and the latitude coefficients from Table 2 (maximum values). It should be noted as well that coefficients from (Romanov, 1962) were assumed as basic. For the cold season, the evapotranspiration was estimated by equations (5) and (6). Then these estimates were averaged for the study basins.

Actual evapotranspiration from the forested area was estimated for each forest type separately (for coniferous and deciduous) and then averaged for the study basins, taking into account weighing coefficients for coniferous and deciduous forests which values were assumed to be proportional to their areas. In this case, to estimate actual evapotranspiration from the forest by equation (4) albedo values \( A_f \) for coniferous and deciduous forests given in Table 1 were used. For spring and fall \( A_f \) values from (Pivovarov, 1977) were used, for winter and summer from (Albedo ..., 1981).

Obtained estimates allowed not only to monitor the changes in the total actual evaporation from the study basins, but also to assess specific contribution to its value of each landscape and land-use elements. Estimations of changes in the total actual evapotranspiration for the Volga and Don river basins and the contribution of each landscape and land-use element into its value are shown in figures 12 and 13. The study period had to be limited by the 1990, as data on saturation deficit used for correction of GGI-3000 pan observations for 20 m² tank data by equation (8) are difficult to obtain for the period of the last 20 years due to Reference-book publication end. At the moment these data are updated. However, even 30-year time series allowed to make some important notions and conclusions.

Figure 12. Temporal changes in the main elements of the water cycle: runoff (1), precipitation (2), and the total evapotranspiration (3) over the Volga (A) and Don (B) river basins.

Despite the smoothed character of fluctuations of the total actual evapotranspiration in the Volga basin, its slight decrease is noted from 1980s, but in the Don basin its negligible changes are observed. At every differences of natural conditions of the two study basins, the results show that interannual fluctuations of evapotranspiration are neutralized by the effect of changes in the areas of landscape and land-use elements. In the Volga basin it is partly forested areas and agricultural land. In the Don basin agricultural land including the area under different crops and irrigated area (Figure 13) are determining factors. Quite unexpected was the relatively small
effect of pan evaporation in the Volga basin, despite the fact that there is a cascade of large reservoirs within this basin. This could mean that in basins with regulated flow determination of annual pan evaporation values by the difference between precipitation and runoff can lead to significant errors.

Figure 13. Relative contribution of various elements of the land cover in the total evaporation over the Volga (A) and Don (B) river basins.

1 - agricultural land: haylands, pastures, multi-year plantations, fallow land;
2 – agricultural crops: spring and winter cereals, industrial crops, potato, vegetables, forage crops;
3 – irrigated crops;
4 – open water surface;
5 – forested land;
6 – swamped land (bogs).

Thus, the estimates of evapotranspiration from the basins of major rivers lead to an important conclusion. It is almost impossible to assess the total actual evapotranspiration from the basin as a whole not taking into account the contribution of specific evaporation from the various landscape and land-use elements and their areas dynamics to its value.

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