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East European river runoff and Black Sea and Caspian Sea level changes as simulated within the Paleoclimate modeling intercomparison project

Alexander Kislov*, Pavel Toropov

Moscow State University, Moscow, Russia

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Abstract

During the last late-Pleistocene and post-glacial epochs, the Caspian Sea fluctuated between regressive and transgressive stages. The Black Sea experienced fluctuations too, but their amplitude was smaller because they were mainly controlled by the World Ocean due to water exchange through the Bosporus Strait. In the water-budget equation, the change in storage over a time period is balanced by the sum of the inflows and outflows that occur during the time period. They can be calculated from atmospheric general circulation models (GCMs). A modeling initiative, the Palaeoclimate Modelling Intercomparison Project (PMIP), could help to resolve this question. The PMIP has focused on two slices of the past: the mid-Holocene (6 ka calendar years Before Present (BP)) and the last cold event of the Late Quaternary (21 ka calendar years BP). The PMIP GCM simulations indicate that changes in level are primarily influenced by changes of river runoff. Ocean precipitation and evaporation together play a minor role. At 21 ka BP the estimated level lowering was ~50 and ~200 m for the Caspian Sea and Black Sea, respectively. This lends credit to the idea of the connection between deep regressive states of Caspian Sea and Black Sea and mature stages of Late Quaternary glacial/cooling/drying planetary events.

1. Introduction

The objects of our exploration are the Caspian Sea and the Black Sea (including the Azov Sea). Fluctuation of their levels is closely related to the main objective of Project IGCP-521. This paper describes a multi-disciplinary investigation, aiming to quantify the relationship between water levels and climate, especially the role of climate in sea level decline. Besides this principal aim, the study of ties between climate change and variations of lake or sea level sand runoff of rivers seems to be very important for a proper interpretation of the future variation of water resources influenced by climate change.

During the last late-Pleistocene and post-glacial epochs, the Caspian Sea fluctuated between regressive and transgressive stages. The Black Sea experienced fluctuations too, but these were mainly controlled by the World Ocean due to water exchange through the Bosporus Strait. Sometimes, the Caspian Sea overflowed to the Black Sea through the Manich Strait (Chepalyga, 2006) and they periodically coalesced. The reasons for change of the seas' levels could be different. This problem can be considered from the point of response of regional-scale water budget to planetary climate changes. The variability of a lake (sea) is a function of balance between the inflow and outflow components of the lake. In the water-budget equation of a lake, the change in storage over a time period is balanced by the sum of the inflows and outflows that occur during the time period. The water-budget components can be expressed in dimensionally consistent volumetric or length units over the time period. All components of the water cycle are functions of climate regime. Therefore, they could be calculated from atmospheric general circulation models (GCMs). Hence, one of the goals of the paper is to discuss how well current GCMs can reproduce river runoff changes, and consequently variations of sea and lake levels in contrasting climate conditions.

A modeling initiative, the Palaeoclimate Modelling Intercomparison Project (PMIP), may help to resolve this question. The goals of PMIP are to compare GCMs driven by the same palaeoclimate forcing in order to understand

*Corresponding author. Tel.: +7 95 939 3043; fax: +7 95 932 8836.
E-mail address: avkislov@mail.ru (A. Kislov).

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better the mechanisms of climate change and test the ability of the models to reproduce climatic conditions radically different from those of today.

The paper is organized as follows. First, previous studies are described (Section 2). Second, global climate changes during the Quaternary and their reasons are described (Section 3). Section 4 contains the PMIP models' descriptions and technique of numerical experiments. Next, the discussion considers the closed-basin lake model that addresses Caspian Sea and Black Sea and restrictions due to data quality (Section 5). Sections 6 and 7 present results from a lake water balance model that address the effects of changes in runoff from the sea's drainage basin and sea levels. Section 8 contains a discussion.

2. Previous studies

Various conceptual models have been used to link climate change to lake level (h), lake surface area (f) and catchment area (F) (e.g., Kalinin, 1968; Street-Perrot and Harrison, 1985; Benson and Paillet, 1989). Kalinin (1968) demonstrated that h asymptotically comes to an equilibrium level that is determined by the steady-state water-budget condition. Street-Perrot and Harrison (1985) classified closed lakes using functions relating f and F based on precipitation and evaporation on the lake surface and catchment (runoff) processes. Benson and Paillet (1989) argued that topographic constraints in lakes with more than one subbasin mean that lake area is the most appropriate measure of lake response to hydrologic balance. Few studies have been able to quantify the lake–climate relationship with precision for individual lakes. Functions of h, f or precipitation and evaporation relationships have been used to evaluate climate model simulations (e.g., Kislov and Sourkova, 1998; Qin et al., 1998) and diagnose past climate (e.g., Harrison et al., 1993, 1996; Jones et al., 2001).

3. Global climate changes during the Quaternary

The Quaternary has been characterized by both cold and warm phases. One candidate for a forcing agent that could produce such pronounced global climate variations is the Milankovitch mechanism (Berger, 1988). According to this theory, the Earth's orbital parameters change due to the influence of the Moon, Sun and planets. Over 100,000- and 400,000-year periods, eccentricity slowly varies, inducing small changes in the annual mean total insolation received by the Earth. Obliquity oscillates from 22° to 25° over a 41,000-year period, and the position of the equinoxes precesses relative to the perihelion with 19,000- and 23,000-year periodicities. Obliquity and precession do not lead to global changes in annual mean energy but strongly modulate the seasonal pattern of insolation.

The evidence for Milankovitch forcing of climate changes may be questioned for several reasons. First, there is no evidence that the climate cycles are periodic rather than aperiodic. Aperiodic factors, such as non-Milankovitch climate variability, could have driven these changes. Core records, both ice and deep-sea, suggest that the dominant character is that of a random red-noise process. Second, much of the energy in low-frequency climate change occurs at periods around 100 ka where the insolation forcing is very weak and the contribution of the Milankovitch frequencies to climate change at most represents only a small fraction (15–20%) of total climate variance. The hypothesis used for description of such type of climate changes can be based on the synthesis of two ideas: (1) climate change as a linear response to Milankovitch's temporal variation of solar insolation; (2) the 'non-linear' hypothesis which extracts the explanations from the theories of stochastical diffusive processes, stochastical resonance and a delayed oscillator model. Non-linear effects mean that orbitally induced insolation forcing is strongly amplified by internal feedbacks. However, sometimes climatic response to variation in insolation can be distinguished from noise at times when orbital agents work synchronously. Such an effect occurred at the transition from the cold late Pleistocene to the warm Holocene. This transition was not gradual; the process was complicated by short-term contingent (possibly random) events (e.g., the Allerød-Younger Dryas-like cycles). Many authors link the origin of these cycles to the behavior of the Atlantic thermohaline circulation (Sarnthein et al., 2000).

4. PMIP models: descriptions

The PMIP (Joussaume and Taylor, 1995) has focused on two slices of the past: the mid-Holocene (6 ka calendar years Before Present (BP) or ~5.3 ka radiocarbon years BP) and the last cold event of the Late Quaternary (21 ka calendar years BP or ~18 ka radiocarbon years BP) because climatic conditions were remarkably different at those times and there are numerous data describing their environmental properties. These slices (together with the modern state) are used in this paper to assess the link between climate variability and hydrologic regime.

A GCM consists of an atmospheric model interactively coupled to submodels of the mixed layer ocean, sea ice, land surface and soil. This model's consideration of terrestrial physical, biophysical and cryospheric processes makes it particularly useful for studying paleoclimate. The atmospheric model has several vertical levels, tending from sigma coordinates near the ground to pressure coordinates at the top of the atmosphere. Models have a typical horizontal resolution of ~2–4° latitude × longitude. The ocean submodel consists of a 50-m-deep thermodynamic slab that represents the ocean mixed layer. Sea surface temperature (SST) and seasonal heat storage vary due to changes in the surface energy balance; however, ocean circulation cannot change. The land-surface model accounts for vegetation effects on near-surface fluxes of momentum, energy and moisture. Typically up to two vegetation layers (upper canopy and lower canopy) can be included at each grid point. The soil submodel controls heat and moisture diffusion through the soil column. It consists of several layers. Surface runoff occurs when the rate of precipitation minus evaporation exceeds the infiltration rate of all soil layers. Water is allowed to drain from the bottom of the soil column; this flux represents near-surface runoff. Designations of the PMIP GCMs and brief descriptions of their spatial resolution are listed in Table 1.

The data from all models were used in this paper except for the following: CLIMBER, MSU and LMD4 because their horizontal resolutions are too coarse.

PMIP GCM experiments are used to understand how climatic parameters important to the Black Sea and Caspian Sea water budget might change due to external forcing. Table 2 specifies all model boundary conditions and parameters.

5. Lake water-budget model and data quality

We calculate changes in lake surface area by solving the steady-state lake water-budget equation at each climatic event, assuming the lake is in hydrologic equilibrium with climate conditions. This is a reasonable assumption when considering the impact of gradual climate change on a lake with a short hydrologic response time compared to the typical time of change of external forcing.

The steady-state equation of the annual budget of water for a closed-basin lake has the form

$$ef = YF \tag{1}$$

where $e = E{-}P$, P is on-lake precipitation m/year per unit lake area, E the lake evaporation m/year per unit lake area, Y the runoff m/year per unit area basin area, F the drainage basin area and f the lake surface area. Eq. (1) assumes that the net groundwater flux into or out of the lake was probably minimal.

Variation of the lake area relative to the present status (denoted by index '0') may be expressed in the form

$$\frac{\Delta f}{f_0} = \frac{\Delta Y}{Y_0} + \frac{\Delta F}{F_0} - \frac{\Delta e}{e_0} \tag{2}$$

It allows evaluating the contribution of different factors to change of the level (h) using information about lake size, bathymetry and the surrounding topography as

$$\Delta h = (\Delta h)_Y + (\Delta h)_F + (\Delta h)_e \tag{3}$$

Evaluation of these factors can be done by different approaches. Information about the change of catchment area $(\Delta h)_F$ can be extracted from paleogeographical data or can be calculated based on simulation of the Earth's surface paleotopography. Level change $(\Delta h)_Y + (\Delta h)_e$ due to both runoff change and change of e is calculated based on data of numerical simulations of GCMs. This method contains an uncertainty or error that reflects errors in all blocks of a GCM. Explicit calculation of $(\Delta h)_Y$ is realized at each numerical time-step of the computer experiment and in each grid point of the land within the GCM. The sum of all values over each basin determines the volume of runoff that occurs during this time period. Another simple approach is a calculation of annually averaged river runoff as a residual in the water-budget equation $Y = P{-}E$ in

Table 1
PMIP models[a] from which data have been used in this paper

Model designation	Resolution[b] and vertical levels (L)
BMRC – Bureau of Meteorology Research Center (Australia)	R21L9
CCC2 – Canadian Center for Climate Modeling and Analysis (Canada)	T32L10
CCM3 – NCAR Climate Community Model (USA)	T42L18
CCSR1 – Center for Climate System Research (Japan)	T21L20
CLIMBER – University of Potsdam (Germany)	7*18L1
CSIRO – Commonwealth Scientific and Industrial Research Organization (Australia)	R21L9
ECHAM3 – Max-Planck Institut für Meteorologie (Germany)	T42L19
GEN2 – National Center for Atmospheric Research (USA)	T31L18
GFDL – Geophysical Fluid Dynamics Laboratory (USA)	R30L20
GISS – Goddard Institute for Space Studies (USA)	72*46L9
LMCELMD4 – Laboratoire de Meteorologie Dynamique (France)	48*36L11
LMCELLMD5 – Laboratoire de Meteorologie Dynamique (France)	64*50L11
MRI2 – Meteorological Research Institute (Japan)	72*46L15
MSU – Moscow State University (Russia)	18*24L3
UGAMP – Universities Global Atmospheric Modeling Programme (UK)	T42L19
UIUC11 – University of Illinois (USA)	72*46L11
UKMO UK – Meteorological Office Unified Model (UK)	96*73L19
YONU – Yonsei University (Korea)	72*46L8

Note: Resolution depends on the type of GCM model: either spectral modes (rhomboidal or triangular) or number of cells determined by latitude × longitude grid points.

[a]These models belong to phase 1 of the PMIP. The second phase of the PMIP has started in 2005.

[b]For example, spectral version T31 provides horizontal resolution ~3.75° latitude × ~3.75° longitude.

Table 2
PMIP boundary conditions and parameters

Boundary conditions and parameters	Control experiment	6 ka BP	21 ka BP
SST and sea ice	Modern	Modern	Calculated or prescribed by CLIMAP
Continental ice sheets	Modern	Modern	Prescribed (Peltier, 1994)
Vegetation and land-surface characteristics	Modern	Modern	Modern (besides areas covered by ice)
Aerosol optical depth	Modern	Modern	Modern
Solar constant ($W\,m^{-2}$)	1365	1365	1365
Orbital parameters	Ecc = 0.016724, $\varepsilon = 23.446°$, $\lambda = 102.04°$	Ecc = 0.018682, $\varepsilon = 24.105°$, $\lambda = 0.87°$	Ecc = 0.018994, $\varepsilon = 22.949°$, $\lambda = 114.42°$
CO_2 (ppm)	280	280	200

each grid cell covering the territory of a catchment. Comparison indicated that the latest method allows achieving more accurate prediction, and therefore it is used in this paper.

Considering today's observed water-budget components of the seas, the volume of water annually delivered by the rivers of the East European Plain (EEP) to the Black Sea is 312 km^3 (total runoff to the Black Sea is 338 km^3). The largest rivers are the Danube, Dneiper, South Bug and Don. Their contribution is ~90% of the total mean volume of the river runoff. The flow of rivers is determined by the balance of precipitation and evaporation; hence, river runoff change responds immediately to climate changes. It is important that both precipitation and evaporation are calculated using a GCM. Moreover, on the EEP, on a large plane, GCM data much better reflect the state of climate compared to areas with complex mosaic conditions of the surface. The contribution of the runoff represents a large fraction (~50%) of inflow volume. Contributions of sea precipitation and inflow water from the Sea of Marmora represent 30 and 20%, respectively. Sea evaporation is 430 km^3 (~53% of outflow volume). Water flux to the Sea of Marmora is responsible for ~47% of the outflow volume.

The Caspian Sea (a vast inland lake) is fed by several rivers. Their annual runoff of the EEP is 274 km^3. The greatest contribution (~80% of the mean total volume of the runoff) is produced by the Volga River. Other principal components of the annual water budget are the precipitation and evaporation over the sea: 76 and 362 km^3, respectively.

A special concern for the climate modeling is the quality of information. Testing based on model-data comparison of monthly (annually) averaged P and E is well established as a method for showing the fitness of a model. Practical realization of this simple idea is complicated by the fact that current GCMs are not able to simulate well regional-scale climate features and peculiarities of the water cycle. This is clearly exhibited when "a point with a point" comparison between modeled and observed data takes place. As a result of this procedure, a wide scatter of data occurs. For example, for the territory of the EEP the range of annual mean temperature anomalies is $\pm 7\,°C$, and the range of relative error of annual precipitation is $\pm 20\%$. This range is evidently decreased by averaging two neighboring points. It decreases by averaging over four points even more, and so on. So, the information at scales involving one or several GCM cells contains large errors. The scale at which the error of modeling data becomes steady and rather small is 1–1.5 million km^2. It is althought that to achieve a good result, it is required to summarize the data over ~15 GCM cells (for the case of typical spatial GCM resolution – see Table 1). Therefore, successful simulation of water balance can be realized for either large rivers only or many rivers simultaneously, by grouping their basins and uniting volumes of runoff.

For the realization of this testing procedure, very good data quality is needed. However, there are inconsistencies over the EEP among various temperature data sets and various precipitation and evaporation data sets (re-analysis of NCEP/NCAR and ECMWF, archives of LEEMENS, GPCP, LEGATES and JAEGER). Preference should not be given to one of them. Therefore, they cannot be used as a reference. Instead of a comparison of modeled and observed P and E, the affinity of the simulated (over the river catchment) and observed (near the mouth of a river) values of the annual river runoff was studied. Large basins covered by many grid cells were used in order to minimize noise that is associated with values from individual grid cells.

Using annual girded values of precipitation and evaporation from each PMIP model, the difference $P-E$ was calculated in each cell of the EEP ($2.5° \times 2.5°$ latitude $\times$ longitude). The data presented in Table 3 denote how well PMIP GCMs simulate today's runoff of the rivers in the basins of the Caspian Sea and Black Sea.

Results were only considered "successful" if the error of modeled volumes of runoff (for each basin) lies within $\pm 20\%$, because the variability does not fall outside the limits of natural variability. According to this

Table 3
Annual volume of runoff water for rivers belonging to basins of Black Sea and Caspian Sea based on data of PMIP GCMs

GCM	Black Sea		Caspian Sea	
	km^3	Error[a]	km^3	Error
BMRC	6.0	−98	68.0	−75
CCC2	182.7	−41	242.6	−11
CCM3	358.2	9	308.7	13
CCSR	341.7	15	195.5	−29
CNRM1	262.1	−16	191.8	−30
CSIRO	251.2	−20	209.5	−24
ECHAM3	161.1	−48	173.6	−37
GEN2	225.2	−27	94.5	−66
GFDL	238.8	−23	44.3	−84
GISS	41.3	−87	94.5	−66
LMD5	653.3	109	187.9	−31
MRI2	233.1	−25	132.8	−52
UGAMP	335.1	7	229.3	−16
UKMO	303.8	−3	138.1	−50
UIUC	515.9	65	141.6	−48
YONU	566.5	82	211.8	−23
Ensemble[b]	292.3	−6	166.5	−39
Successful[c]	308.7	−1	260.2	−5
Observed[d]	312	274		

[a]Relative error (%).
[b]Data of the ensemble of all PMIP GCMs.
[c]Data of the ensemble of "successful" PMIP GCMs.
[d]Only rivers from the EEP.

classification, the "successful" GCMs were chosen. The data of these models were examined more closely under modeling of other climatic regimes. Comparison with the observed data has shown that a few models are capable of reproducing runoff to the Black Sea and the Caspian Sea. The direct reason is that annual precipitation in steppe zones is essentially underestimated by the majority of GCMs.

6. Simulation of river runoff and sea level changes during the mid-Holocene warm event

The Holocene is marked by a relatively stable warm climate experiencing weak changes on a global average. By the Early Holocene (~9 ka BP), CO_2 was up to its pre-industrial level (Raynaud et al., 1993), the Scandinavian ice sheet was almost gone and the Laurentide ice sheet had shrunk considerably (Dyke and Prest, 1987; Svensson, 1991). SST was not significantly different from today. In the mid-Holocene, at 6 ka BP, insolation anomalies were +5% in summer and −5% in winter, less than those at the beginning of the Holocene. PMIP simulations of the 6 ka BP climate were thus designed as pure sensitivity experiments to changes of insolation forcing (see Table 2).

All PMIP models simulate an increased seasonal cycle of temperature over the continents in the Northern Hemisphere, reaching ~1 °C on global average for summer/winter seasons. In summer, all models demonstrate a temperature increase over land in the temperate zone. All models simulate an increase of African and Asian monsoon rains in response to additional summer warming of the land (Joussaume et al., 1999). Global model results have been evaluated (Joussaume, 1999; Kohfeld and Harrison, 2000) against terrestrial proxy data (lake levels, pollen- and plant macrofossil-based reconstructions of 6 ka BP vegetation). Comparing these paleoenvironmental estimates of temperature and precipitation changes at 6 ka BP with modeling data indicates their close agreement but shows that modeled precipitation is underestimated in several arid and semi-arid regions. Model-data discrepancies may result in part from the omission of vegetation and ocean temperature feedback on climate (Joussaume, 1999; Otto-Bliesner, 1999; Braconnot et al., 2000).

The EEP winter mean cooling is obtained by all models forced by the orbitally induced negative anomalies of insolation; however, the simulated changes are very small. The patterns of spatial distribution of spring and autumn mean temperature are essentially similar to today. The models show winter conditions slightly drier than today (5%) in the central, northern and eastern parts of the region, whereas conditions drier than today (10%) form in the western part of the EEP. At 6 ka BP, the models show insolation-induced summer conditions in the EEP that were warmer (1–1.5 °C) than today. Maximum summer mean warming 2 °C occurred in the southwestern part of the territory. PMIP simulation of the 6 ka BP summer shows conditions drier than today over practically the whole EEP.

Calculation indicates that at 6 ka BP there is no significant change of the EEP river runoff (see Table 4). The Volga River contribution to the water running to the Caspian Sea slightly increases (93%) compared to today (88% due to PMIP models' simulation and 84% based on observation). The value $\Delta e/e_0$ over the Caspian Sea was estimated (based on regional climate modeling (Kislov and Sourkova, 1998)) as a small value relative to the first term in Eq. (2).

In spite of the slight response of river runoff to external orbital forcing, it is unexpected that during the Holocene the Caspian Sea level was not stable. The amplitude of perturbations has a value of several meters. Such changes could be produced by non-climatic factors such as changes in the inflows to or outflows from the lake due to tectonic processes, but more probably they reflect the extent of the inaccuracy of the estimation of water-budget components.

Table 4
Change of runoff of rivers ($(Y-Y_0)/Y_0$) (%) belonging to Black Sea and Caspian Sea at 6 ka BP based on PMIP data

	Black Sea	Caspian Sea
Ensemble of PMIP models	+9	+14
The successful PMIP models[a]	−5	+5

[a]They are CCM3, CCSR, CNRM1, UGAMP and UKMO for the Black Sea and CCC2, CCM3 and UGAMP for the Caspian Sea.

7. Simulation of river runoff and sea level changes during the last cold Pleistocene event

The last cold event of the Pleistocene involved substantial changes in surface boundary conditions, and within the PMIP their specific set was prepared (see Table 2). At ~21 ka BP, the latitudinal distribution of insolation and its relative seasonal strength were similar to today (for example, in the northern summer, the solar energy deficit received by the Earth was ~2–4 W m^{-2}). The ice sheet extent and height have been provided by Peltier (1994). CO_2 concentration was estimated at 200 ppm as inferred from Antarctic ice cores (Raynaud et al., 1993). Over the oceans, two sets of PMIP experiments were defined: (1) prescribing SST changes from estimates given by CLIMAP (1981) and (2) computing SST using coupled atmosphere-mixed layer ocean models. Clearly, these experiments show only a limited range of climatic variations that are affected by selected changes. Other kinds of intrinsic variability (e.g., due to ocean circulation changes) or other kinds of natural forcing (e.g., solar irradiance and volcanic forcing) are not encompassed by these experiments.

Note that the boundary conditions used by the PMIP contain some uncertainties. For example, the location of ice sheets reconstructed by Grosswald (1998) has been extended over all parts of the Arctic shelf to cover large portions of the continents, but from the point of view of Vasil'chuk and Vasil'chuk (1995), the "glacial maximum" is expressed by an increase in mountain glaciation and spreading of permafrost conditions. Another example touches upon an underestimation of CLIMAP SST anomalies in the tropics (Anderson and Webb, 1994; Guilderson et al., 1994; Hostetler and Mix, 1999) and the problem of location and seasonal behavior of sea-ice cover (Weinelt et al., 1996; Sarnthein et al., 2000).

Global model results have been evaluated against terrestrial proxy data (Joussaume, 1999; Kohfeld and Harrison, 2000; Kislov et al., 2002). These results clearly demonstrate that temperatures modeled in the PMIP experiment reproduce the main peculiarities of reconstructed land temperature fields, but over the tropics, the simulations with prescribed CLIMAP SSTs produce too weak a cooling effect over land. All models produce drying in the extra-tropical zone, although the extent and location of the regions of increased aridity vary between models. Comparing the results of two sets of experiments with prescribed and computed SSTs indicates that the prescribed SSTs should yield a better model-data fit over land. This is an interesting fact, because there is no a priori reason to expect this to be so given the uncertainties in SST reconstructions.

Using PMIP data, we now calculate ensemble mean and standard intermodel deviation values for the EEP. Winter mean cooling is obtained in the EEP by all PMIP models. Maximum cooling 20 °C occurs in the northern part of the EEP. Over other parts of the EEP, the negative anomaly ~10 °C is obtained. The spatial distribution of the intermodel standard deviation is much more uniform and substantially less than 10 °C. This finding means that the temperature patterns simulated by the different models have many principal details in common.

Concerning uncertainties in the data input into the models, it is noted that the underestimation of CLIMAP SST anomalies in the tropics produces, probably, too weak a simulated cooling effect over land of tropical belt. Model sensitivity uncertainties related to the glaciological information have been investigated previously (Kislov et al., 2002). Results of three simulations were compared where different glaciological scenarios were used as boundary conditions: extremely large arctic ice sheet; extremely small sheet (but this scenario includes widespread permafrost conditions); and a traditional CLIMAP scenario. It was shown that most differences (2–6 °C) between the results occurred within the regions where the position of the ice sheet has been changed. So, the difference in boundary condition provides only a regional effect. Moreover, taking into account the intermodel deviations of simulation data, these uncertainties are not crucial for global climate modeling.

Consider the results of calculation of annual river runoff volumes for the Caspian Sea and the Black Sea (Table 5). At 21 ka BP the total river runoff to the Caspian Sea (calculated by "successful" models) is substantially decreased (~50%) compared to today. The relative contribution of Volga River runoff has a value 72%. These facts are in accordance with the observational data.

Information about river runoff change provides a useful guide for conclusions about the status of the Caspian Sea and the Black Sea. It is supposed (based on geomorphological evidence) that configuration of the catchment area of the rivers has not been principally changed at 21 ka BP; therefore, the second term in Eq. (3) is equal to zero. Annual lake evaporation is shown to be higher for warming conditions (e.g., Morrill et al., 2001), and therefore levels could be increased at 21 ka BP due to this effect, but annual precipitation over the sea was substantially smaller than today as demonstrated by the PMIP GCMs (Kislov and Toropov, 2006). The value $\Delta e/e_0$ can be estimated as a small value relative to the first term in Eq. (2). Using Eq. (2), the relative decrease of the Caspian Sea area is 50%, and it means (using a relationship between Δf and Δh) a substantial dropping of level (~50 m) (see Fig. 1).

Table 5
Change of runoff of rivers ($(Y-Y_0)/Y_0$) (%) belonging to Black Sea and Caspian Sea at 21 ka BP based on PMIP data

	Black Sea	Caspian Sea
Ensemble of PMIP models	−22	−40
The successful PMIP models	−45	−56

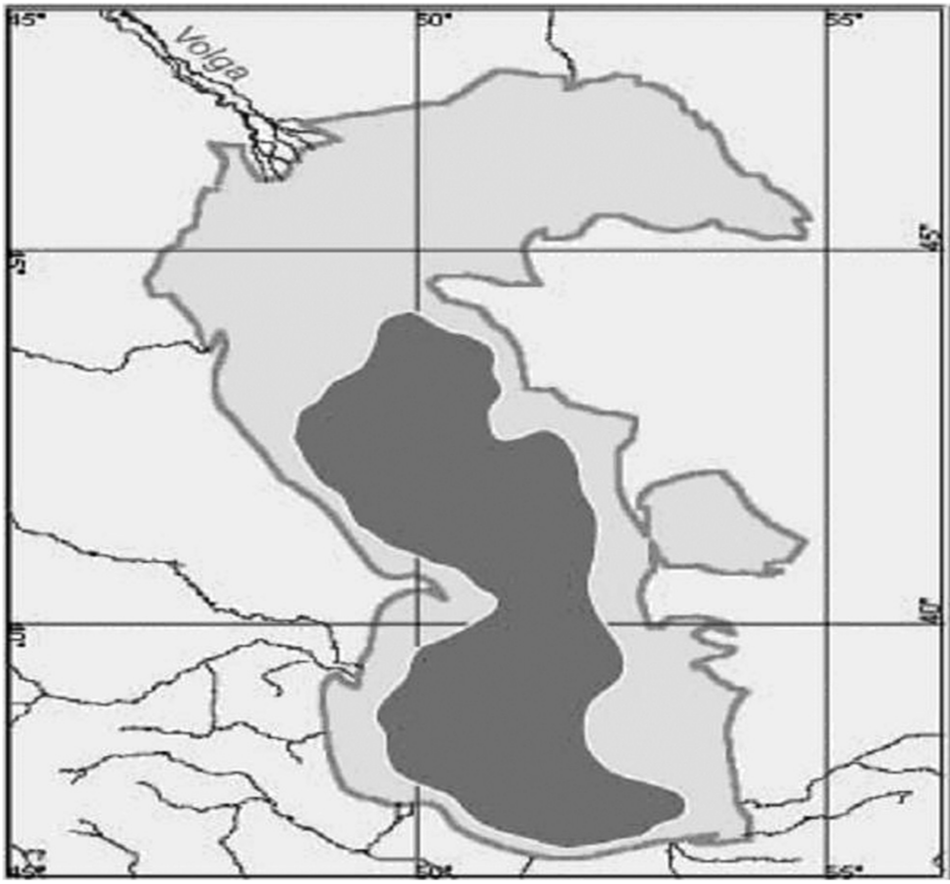

Fig. 1.

As far as the Black Sea is concerned, its calculated river runoff is substantially decreased (~45%) as well. This fact, coupled with an assumption that due to decreasing level the Black Sea at 21 ka BP was a closed lake, allowed the use of Eqs. (2) and (3) and an estimate that the drop of level was ~200 m (Fig. 2). During this period, there was no water exchange through the modern Bosporus Strait due to dropping of both the Black Sea and the World Ocean.

8. Discussion

A simple water balance model was presented, and the ability of the PMIP GCMs to simulate changes in the level of the closed-basin sea has been demonstrated. Processes by which orbital forcing and glacial external forcing may have altered the water balance of Black Sea and Caspian Sea were examined. The goals were to provide more accurate paleoclimate interpretations of water level fluctuations and estimate the hydrologic response of these seas to climate change. GCM simulations indicate that level changes are primarily influenced by changes of river runoff. Sea precipitation and sea evaporation together play a minor role. The global model does not provide a perfect representation of all processes, however. It is needed to couple the atmospheric model to the ocean model, take into account climate–vegetation feedback and improve boundary conditions. On a regional scale, there are several processes that influence runoff that the GCM does not either include or represent well. First, vegetation does not vary in the GCM between the three studied cases. Second, change of permafrost is not taken into account. Such effects are evaluated to induce changes as large as 0.3 in the runoff coefficient, i.e., the ratio of runoff to precipitation. Although the results in these areas are still speculative due to a lack of field data, it appears plausible that changes in the runoff coefficient could impact lake level change.

These results are important in the light of a problem of chronology of paleogeographical events belonging to different regional scales. In response to glacial external forcing conditions, lowering levels of the Black Sea and Caspian Sea are simulated simultaneously. These events reflect the so-called Atelskya regression stage of the Caspian Sea and the so-called Postkarangatskya regression stage of the Black Sea. Hence, these changes cannot be due to local climate forcing: they probably reflect planetary-scale climate forcing. This lends credit to the idea of the connection between deep regression states of the Caspian Sea and Black Sea and mature stages of Late Quaternary glacial/cooling/drying planetary events.

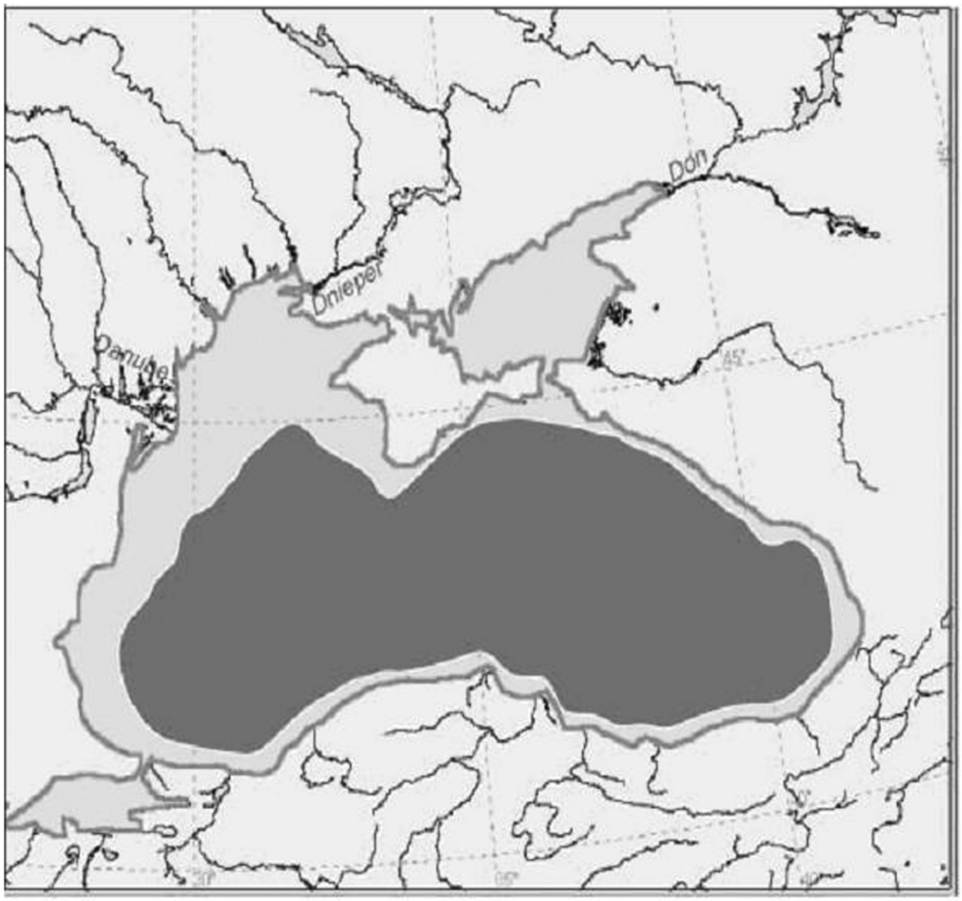

Fig. 2.

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