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Simulating long-term Caspian Sea level changes: The impact of Holocene and future climate conditions

H. Renssen [a,*], B.C. Lougheed [a], J.C.J.H. Aerts [b], H. de Moel [b], P.J. Ward [a], J.C.J. Kwadijk [c]

[a] *Institute of Earth Sciences, Faculty of Earth and Life Sciences, Vrije Universiteit Amsterdam, De Boelelaan 1085, NL-1081HV Amsterdam, The Netherlands*
[b] *Institute for Environmental Studies, Faculty of Earth and Life Sciences, Vrije Universiteit Amsterdam, De Boelelaan 1085, NL-1081HV Amsterdam, The Netherlands*
[c] *WL-Delft Hydraulics, P.O. Box 177, NL-2600MH Delft, The Netherlands*

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Abstract

To improve our understanding of the relationship between climate change and variations in Caspian Sea level (CSL), we performed simulations of annual CSL for the period 8 ka to 2100 CE using a coupled model setup representing climate, hydrology and sea level. We forced our climate model with long-term changes in orbital parameters and atmospheric greenhouse gas concentrations, using the IPCC A1b scenario for the 21st Century. Our simulations produce an orbitally forced, long-term decline in CSL of 5 m from 5.5 to 0 ka, caused by a decrease in river runoff and over-sea precipitation that is not fully compensated by a decrease in over-sea evaporation. Superimposed on this long-term downward CSL trend we simulated centennial-scale fluctuations of up to 4 m and decadal-scale variations of up to 2 m, caused by the internal variations of our modeled climate system, amplified by the sensitivity of CSL to small changes in river runoff and in the over-sea P–E budget. The A1b anthropogenic emission scenario causes a 4.5 m fall in CSL in the 21st Century, due to a pronounced increase in over-sea evaporation that is stronger than the enhanced river discharge. This decline in CSL is of the same order of magnitude as the orbitally-forced millennial-scale downward CSL trend simulated for the last 8000 years. Our results are generally consistent with CSL estimates based on geological, historical and measured data, as well as with most other model studies.

Keywords: Caspian Sea; sea level; Volga; Holocene; model simulations; future climate

1. Introduction

The Caspian Sea level (CSL) has experienced substantial fluctuations during the 20th Century. In the 1930s, the level fell abruptly by about 2 m, whilst a rapid rise of the same order of magnitude was observed after 1977 (Rodionov, 1994). These decadal-scale variations are superimposed on a long-term downward trend in CSL, as suggested by geological evidence derived from the dating of deltas and terraces (i.e. CSL high stands, Rychagov, 1997). This geological evidence shows a decrease of about 5 m from a level of −20 mbsl (meter below sea level) in the early Holocene (around

* Corresponding author. Tel.: +31 20 5987376; fax: +31 20 5989941.
E-mail address: hans.renssen@falw.vu.nl (H. Renssen).

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8 ka) to a mean level of −25 mbsl in the 18th–19th Centuries CE (Rychagov, 1997). It is generally accepted that climate-induced changes in the hydrological budget of the Caspian Sea are the main cause of such CSL fluctuations (Rodionov, 1994; Elguindi and Giorgi, 2006a,b). However, as noted by Rodionov (1994), geological processes are also thought to influence the CSL, including tectonic movements (Vdovykin, 1990) and deep groundwater flows between the Aral Sea and the Caspian Sea (Shilo, 1989). In addition, anthropogenic activities such as land-use change and reservoir development have affected the CSL during the 20th Century (Rodionov, 1994).

In view of ongoing global warming, it is important to gain a thorough understanding of the relationship between climate and CSL. Recently, several modeling studies have been performed that shed light on this relationship. Elguindi and Giorgi (2006a) were able to reproduce the decadal-scale CSL variations observed during the 20th Century, using a regional climate model for the Caspian Sea basin combined with a simple hydrological water-balance equation. The same model was applied to estimate the future CSL evolution following the A2 IPCC (2001) scenario, suggesting a large CSL decrease of about 5 m by the end of the 21st Century due to increased evaporation loss (Elguindi and Giorgi, 2007). However, a much more ambiguous picture of the future emerged from an analysis of CSL variations using seven different Atmosphere–Ocean general circulation models (AOGCMs) forced by the same A2 scenario (Elguindi and Giorgi, 2006b). Some of these models produced a CSL drop of more than 10 m by the end of the 21st Century, while other models suggested a stable or even increasing CSL (Elguindi and Giorgi, 2006b; Arpe and Leroy, in press).

The uncertainty about future CSL shows that it is necessary to improve our understanding of the mechanism behind CSL variations. One way forward is to analyze long-term variations that have occurred during times predating significant anthropogenic influences, and to compare these 'natural' fluctuations with projections of future CSL changes. We have therefore simulated the variations in CSL for the period 8 ka until 2100 CE, using a hydrological model that is specifically setup for the Caspian Sea basin in combination with a simple model for the CSL. The climatic data used to force our hydrological model are derived from a transient climate model simulation of the same period, performed by Renssen et al. (2005), extended with an IPCC A1b scenario run for the 21st Century. Compared to earlier studies of past CSL variations, our transient coupled climate-hydrological model approach is novel,

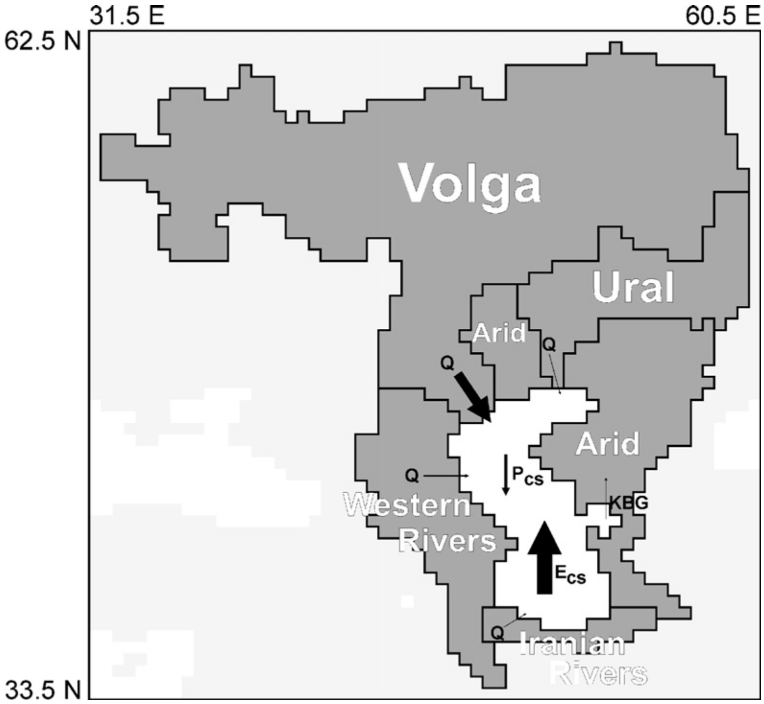

Fig. 1. Overview of the Caspian Sea drainage basin on the STREAM grid, showing the catchments of the main contributing rivers (Volga, Ural, Iranian Rivers and rivers to the West of the Caspian Sea). Also shown are the principal elements of Eq. (1), i.e. river runoff (Q), over-sea precipitation (P_{CS}) and evaporation (E_{CS}), and flow to Bay of Kara–Bogaz–Gol (KBG).

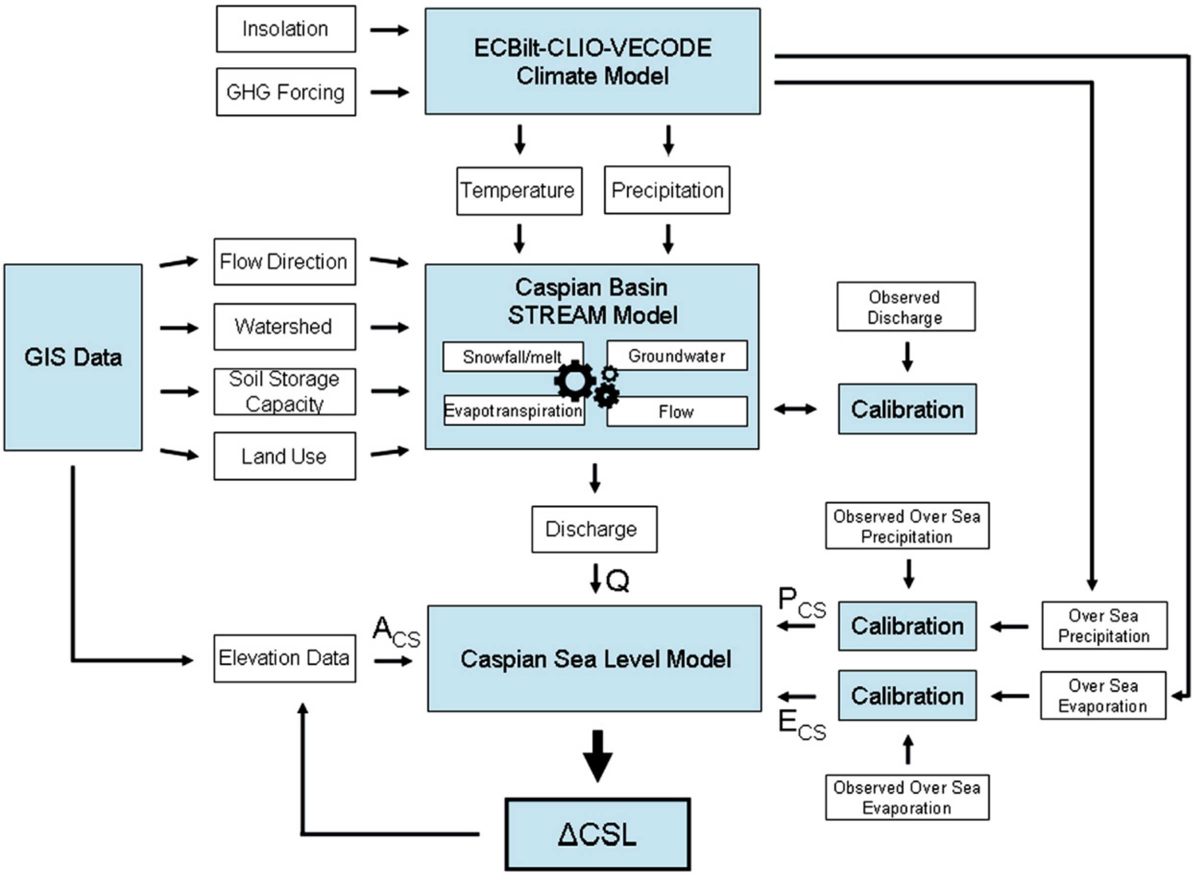

Fig. 2. Flowchart of our coupled model, showing the different components (see text for details).

as these studies have used only climate model output for specific time slices (e.g., 6 ka, Kislov and Surkova, 1998; Kislov and Toropov, 2006, 2007). In this paper we address the following specific research questions:

- How much of the reconstructed long-term (i.e. millennial-scale) variations in CSL can be attributed to a response to slowly changing climate forcings?
- What is the internal variability of the coupled system on decadal-centennial time-scales over the course of the Holocene?
- How do expected future changes in CSL compare to these long-term changes and to internal variability?
- Is the warm early Holocene period a suitable analogue for the future?

2. Methods

We have setup a coupled model for our CSL simulations, representing climate, hydrology and sea level (Figs. 1 and 2). We have estimated annual variations in CSL using the following simple equation adapted from Rodionov (1994):

$$\Delta L = Q/A_{\mathrm{CS}} + P_{\mathrm{CS}} - E_{\mathrm{CS}} - KBG \qquad (1)$$

where ΔL is change in CSL (cm/yr), Q is total river discharge (cm^3/yr), A_{CS} is Caspian Sea surface area (cm^2), P_{CS} is over-sea precipitation (cm/yr), E_{CS} is over-sea evaporation (cm/yr) and KBG is the sea-level loss due to flow to the Bay of Kara–Bogaz–Gol (cm/yr) (Fig. 1). This bay is a separate water body that is connected by a 250 m wide channel to the eastern part of the central Caspian Sea. The annual loss to the Bay of Kara–Bogaz–Gol was fixed at 2.7 cm/yr following Rodionov (1994). The surface area of the Caspian Sea (A_{CS}) experiences a relatively large change with varying CSL because of the very gentle slope of the northern shore. We varied A_{CS} with CSL according to an empirical linear relationship based on a digital elevation model (Lougheed, 2006):

$$A_{\mathrm{CS}} = 8305.7CSL + 608239. \qquad (2)$$

The total river discharge (Q) was simulated with a hydrological model setup for the Caspian Sea following the principles of the STREAM model, described in detail in previous studies (e.g., Kwadijk, 1993; Aerts et al., 1999; Ward et al., 2007). The model was run at a resolution of 0.5° by 0.5° (Fig. 1) and with a monthly time step. The impact of anthropogenic water extraction or other artificial constructions is not accounted for. The model simulates the total runoff entering the Caspian Sea. However, at this spatial resolution only large river basins can be delineated accurately. The Volga is the largest supplying basin and the most important contributor to Q (Meshcherskaya and Golod, 2003), producing about 82% of the total river runoff entering the Caspian Sea (Shiklomanov et al., 1995). STREAM was therefore calibrated on monthly Volga discharge data at Volgograd for the period 1903–1935 (Aerts et al., 2006; Lougheed, 2006), i.e. before major anthropogenic influences on the river (Rodionov, 1994). This calibration was successful, yielding a Pearson product-moment correlation coefficient of $R=0.964$ between observed and modeled Volga runoff data (Lougheed, 2006). The other contributing rivers (e.g. Ural, Sulak, Samur, Kura and Terek) have smaller basins and are more difficult to represent accurately. Therefore the simulated discharge of these rivers was calibrated such that the ratio between the discharge of the other rivers and the Volga was correct for the 20th Century (Lougheed, 2006). This calibration factor was also used for the simulated period between 8 ka and 2100 CE.

Next, the calibrated STREAM model was forced with monthly precipitation and temperature model results from a transient simulation for the period 8 ka to 2100 CE performed with version 3 of the ECBilt-CLIO-VECODE coupled atmosphere–ocean–vegetation model (Opsteegh et al., 1998; Goosse and Fichefet, 1999; Brovkin et al., 2002). As the latter model simulates the global climate at a relatively coarse spatial resolution of 5.6° latitude by 5.6° longitude, the climatic data were spatially downscaled and redistributed following Aerts et al. (2006) and Ward et al. (2007). For the period 8 ka to 1750 CE, ECBilt-CLIO-VECODE was forced by long-term annual variations in orbital parameters and atmospheric CO_2 and CH_4 concentrations (Renssen et al., 2005), whilst for the period 1750 to 2100 CE the prescribed rise in greenhouse gas levels followed the anthropogenic SRES A1b emission scenario (IPCC, 2001). This scenario was quantified using the IMAGE-2 model (IMAGE team, 2001), and results in an equivalent CO_2 concentration of 1050 ppmv in 2100 CE. Relative to other SRES emission scenarios, A1b represents an intermediate greenhouse gas forcing for the 21st Century. For the anthropogenic period (1750–2100 CE) the increasing greenhouse gas concentration is the dominant climate forcing mechanism, whilst for the pre-anthropogenic period (8 ka–1750 CE) orbital forcing is more important given the relatively small changes in CO_2 and CH_4 (Renssen et al., 2005).

Over-sea precipitation (P_{CS}) and evaporation (E_{CS}) were derived from the same simulation of ECBilt-CLIO-VECODE (Renssen et al., 2005). However, as the low resolution land–sea mask of this climate model does not resolve the Caspian Sea, we were not able to use the simulated local precipitation and evaporation results for P_{CS} and E_{CS}, respectively. Instead, we utilized the precipitation and evaporation values of the most eastern Mediterranean Sea cell, which were linearly converted to resemble values over the Caspian Sea using the equation of Bouwer et al. (2004). As detailed by Lougheed (2006), this equation was used to adjust the standard deviation and mean of the Mediterranean ECBilt-CLIO-VECODE P and E values, with conversion factors derived using precipitation and evaporation estimates for the Caspian Sea over the period 1901–2000 CE, provided by Baidin and Kosarev (1986) and Rodionov (1994). As a result of this conversion, our mean P_{CS} and E_{CS} values for 1901–2000 CE and their standard deviations closely resemble the actual over-sea estimates for the same period. In this study, we assume that the conversion factors are valid for the entire period 8 ka to 2100 CE. Regrettably, we are unable to test this assumption, as estimates for P_{CS} and E_{CS} are unavailable for periods before the 20th Century. Potentially, past changes in the large-scale atmospheric circulation may have resulted in a different relationship between precipitation and evaporation over the Eastern Mediterranean and Caspian Seas.

To evaluate the performance of our coupled model setup (Fig. 2), Lougheed (2006) compared the simulated CSL variations for the 20th Century with observed values of Rodionov (1994). This comparison revealed that our model captures the 20th Century CSL reasonably well, producing values between −27.0 and −25.0 mbsl, compared to a reconstructed range between −28.0 and −25.9 mbsl; the latter estimate was corrected for anthropogenic disturbance (Rodionov, 1994). Consequently, we assume that our model is an appropriate tool for analyzing the response of CSL to climate change.

Our calculations are not very sensitive to the initial CSL value used for 8 ka. We have performed tests with different initial CSL values ranging between −15 mbsl and −30 mbsl, using the same input in Eq. (1) for the period 8 to 7 ka. In these tests, all results converged to a CSL of −21.5 mbsl within 300 years. We have therefore used −21.5 mbsl as an initial CSL value.

3. Results and discussion

3.1. Long-term variations

Over the course of the last 8000 years, our model suggests a long-term decrease in CSL, with a total drop of 5 m prior to the start of the industrial era (Fig. 3). Between 8 and 5.5 ka, levels are still relatively stable with CSL values fluctuating around −21.5 mbsl. After 5.5 ka, a long-term CSL decrease of about 1 m per millennium sets in until a level of −26.5 mbsl is reached in the 17th Century CE. The trend after 5.5 ka is not linear, as the decline in CSL accelerates slightly over time. This long-term CSL trend is consistent with available geological evidence for sea level high stands, also suggesting a gradual millennial-scale decline in CSL of about 5 m, from −20 mbsl just after 8 ka, to −21 mbsl at 6.8 ka, −23 mbsl around 3 ka and −25 mbsl in the 18th–19th Centuries CE (Fig. 3, Rychagov, 1997). Rychagov (1997) infers that several long-lasting low stands occurred between these high stands (e.g., between 5.5 and 4.5 ka, and the so-called 'Derbent regression' between 2 and 1 ka), with CSL values as low as −30 to −32 mbsl. However, it should be noted that the geological evidence for these low stands is much more uncertain than for the high stands (Rychagov, 1997; Hoogendoorn et al., 2005; Kroonenberg et al., in press). In any case, our simulations do not reproduce these long-lasting low stands. In addition, most of the radiometric ages on which the CSL curve of Rychagov (1997) is based are uncertain, as they are derived from 30-year old bulk ^{14}C analyses on large samples of molluscs (Kroonenberg et al., in press). Unfortunately, the CSL curve of Rychagov (1997) is the only one available.

The simulated millennial-scale CSL decrease is caused by a delicate interplay between all contributing factors (Q/A_{CS}, P_{CS} and E_{CS}, Fig. 4). All these factors experience a long-term decrease from the early part of the run (8–5.5 ka) until the last millennium: Q/A_{CS} by 0.4 cm/yr, P_{CS} by 3.2 cm/yr and E_{CS} by 3.5 cm/yr (Table 1). As the decrease in input of 3.6 cm/yr ($Q/A_{CS}+P_{CS}$) is slightly larger (by 0.1 cm/yr) than the decrease in output (E_{CS}), the CSL falls by about 5 m over the last 5000 years.

The long-term CSL decline follows the orbitally-forced decrease in insolation during the summer season (June to September, or JJAS), which is reduced by about 16 W/m^2 in the last 8000 years at 50°N (Fig. 3). This JJAS insolation trend is similar in shape to the simulated CSL trend, showing the largest decrease after 5 ka. As a result of higher summer insolation values, early Holocene summers were warmer and both evaporation and precipitation rates were higher on mid-latitude continents in the Northern Hemisphere (Renssen et al., 2005).

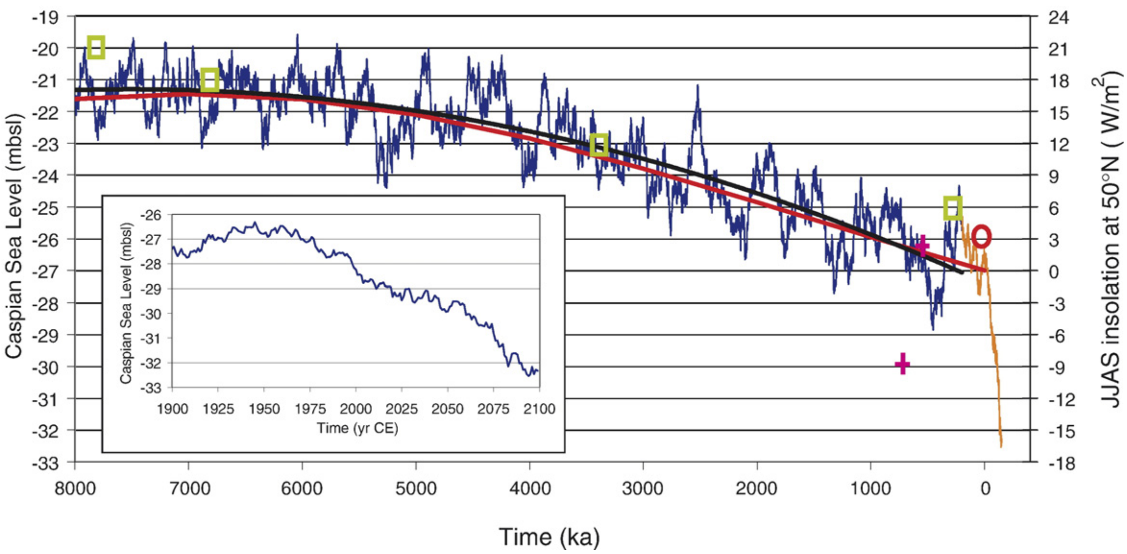

Fig. 3. Simulated annual CSL evolution (left axis), with in blue the period 8 ka to 1750 CE, and in orange the period 1750 to 2100 CE. The black line is a 2nd order polynomial fit applied to the CSL data for 8 ka–1750 CE, representing the long-term orbitally-forced trend. Also shown in red is the change in JJAS insolation at 50°N, i.e. the main climatic forcing (right axis, relative to the 0 ka value). Green boxes indicate the CSL estimates based on geological evidence (Rychagov, 1997), purple crosses indicate estimates based on historical data (Rodionov, 1994) and the red circle is the mean measured CSL for the early 20th Century. The inset shows the simulated CSL curve for the 20th and 21st Centuries. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

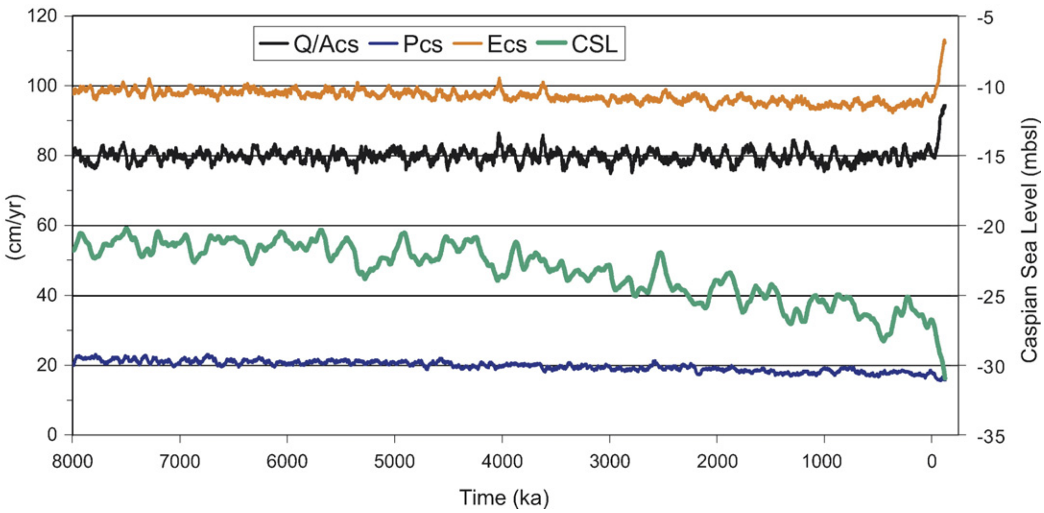

Fig. 4. Simulated long-term evolution of factors contributing to CSL (50-year running means, left axis): total river discharge divided by the Caspian Sea surface area (Q/A_{CS}, black), over-sea precipitation (P_{CS}, blue) and over-sea evaporation (E_{CS}, orange). Also shown is the 50-year running mean CSL for comparison (green, right axis). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

3.2. Decadal-centennial variability

Superimposed on the simulated long-term millennial-scale decline in CSL are numerous fluctuations on decadal-to-centennial time-scales (Fig. 3). These fluctuations must originate from the internal variability of our coupled atmosphere–ocean–vegetation system. The other components (STREAM and sea-level equations) do not incorporate internal fluctuations and the external climate forcings that were prescribed (i.e. orbital and greenhouse gas forcing) in the ECBilt-CLIO-VECODE simulation have no variability on these time-scales (Renssen et al., 2005). Indeed, experiments performed with our coupled atmosphere–ocean model (ECBilt-CLIO) with constant forcings have been shown to exhibit significant variability at various time-scales, including decadal variability that resembles the North Atlantic Oscillation (e.g., Goosse et al., 2001, 2002, 2003).

Major fluctuations in CSL occur when river runoff and $P_{CS}-E_{CS}$ operate in the same direction. In other words, important drops in CSL take place when low runoff values coincide with major reductions in $P_{CS}-E_{CS}$, and vice versa. Consequently, subtle internal variations of the climate system that are expressed in small fluctuations in precipitation and evaporation are greatly amplified by the CSL. This is due to the sensitivity of CSL to slight changes in $P_{CS}-E_{CS}$ and inflow from contributing rivers. Potentially, this could also be the case for other large inland lake systems.

Table 1
Summary of mean values calculated for each millennium and for the 20th and 21st Centuries CE

Time	Total river discharge Q (km^3/yr)	Volga discharge (km^3/yr)	"Other" discharge (km^3/yr)	A_{CS} (km^2)	Q/A_{CS} (cm/yr)	P_{CS} (cm/yr)	E_{CS} (cm/yr)	$P_{CS}-E_{CS}$ (cm/yr)	ΔL (cm/yr)	Mean CSL (mbsl) with (SD)
8–7 ka	343.7	272.7	71.0	431030	79.8	21.5	98.5	−77.0	0.06	−21.3 (0.7)
7–6 ka	343.8	269.1	74.6	430414	79.9	21.1	98.3	−77.2	0.03	−21.4 (0.7)
6–5 ka	339.4	266.7	72.7	425757	79.7	20.9	98.0	−77.1	−0.04	−22.0 (0.9)
5–4 ka	340.6	265.7	74.9	426099	80.0	20.2	97.7	−77.5	−0.25	−21.9 (0.9)
4–3 ka	333.4	262.1	71.3	417659	79.8	19.7	96.7	−77.0	0.14	−22.9 (0.7)
3–2 ka	322.5	256.1	66.4	405794	79.5	19.1	96.0	−76.9	−0.11	−24.4 (0.9)
2–1 ka	317.5	255.9	61.6	398711	79.7	18.2	95.3	−77.1	−0.19	−25.2 (0.9)
1–0 ka	309.3	250.4	58.9	389731	79.4	18.0	94.8	−76.8	−0.14	−26.3 (0.9)
20th Cent. CE	309.8	252.5	57.3	383225	80.8	17.7	96.5	−78.8	−0.65	−27.1
21st Cent. CE	328.0	269.2	58.8	359130	91.4	15.7	108.6	−92.9	−4.16	−30.0

For the CSL the standard deviations (between brackets) were calculated after removal of the long-term cooling trend depicted in Fig. 3.

Generally, two types of CSL fluctuations can be observed, differing in time-scale and magnitude. First, centennial-scale fluctuations occur with a maximum magnitude of about 4 m (e.g., 2.7–2.5 ka, Fig. 3). Second, decadal-scale variations occur with a maximum extent of about 2 m. The amplitude of this decadal–centennial-scale variability increases after 6 ka, as indicated by the CSL standard deviations that were calculated per millennium after removal of the long-term trend (Table 1). The magnitude and temporal scale of these simulated CSL fluctuations is consistent with historical evidence from the last millennium and with 20th Century measurements. According to Rodionov (1994), CSL experienced a strong increase of about 5 m during the 13th–14th Centuries CE (Fig. 3), which could be an example of the noted centennial-scale CSL variations. Likewise, the measured 2 m drop in the 1930 s is similar to the simulated decadal-scale fluctuations. It is important to note that we cannot expect that the decadal–centennial-scale fluctuations generated by our models have a similar timing as those that occurred in the real world. A first reason is that our climate model experiment represents only one out of many valid realizations of climate. A second reason is that we have not considered the role of short-term forcings such as solar variability and volcanic eruptions, which could potentially also have influenced CSL variability on these time-scales.

3.3. 20th–21st Centuries warming

In the 21st Century, the simulated CSL shows a marked decline of 4.2 m from −28.3 to −32.5 mbsl as a result of the anthropogenic A1b scenario (Fig. 3). In our model, this decline already starts in the 1970 s (CSL of −26.5 mbsl), when temperatures begin to rise due to the steady increase in greenhouse gases, producing a total CSL fall of 6.0 m for the period 1975–2100. As a result of this warming, over-sea evaporation increased considerably by 12.1 cm/yr from the 20th to the 21st Century, overwhelming the increase in river runoff of 10.6 cm/yr (Table 1, Fig. 4). The increase in total runoff is mainly due to enhanced 21st Century precipitation over the Volga catchment. Over-sea precipitation slightly decreased from the 20th to the 21st Century (by 2.0 cm/yr), thus contributing to the decrease in CSL. Our result is consistent with Elguindi and Giorgi (2006b), who analyzed the projected 21st Century CSL change using the same A1b scenario in seven AOGCMs. Six out of seven models predict a fall in CSL, with their ensemble mean suggesting a drop of about 7 m from −27 mbsl in 2000 CE to −34 mbsl in 2100 CE. In these models, the enhanced evaporative loss in the 21st Century also overwhelms the increase in Volga runoff (Elguindi and Giorgi, 2006b). However, in other AOGCM A1b scenario runs the enhanced Volga runoff dominates, leading to an increase in CSL in the 21st Century (Elguindi and Giorgi, 2006b; Arpe and Leroy, in press). It must be stressed once again that we have not included direct anthropogenic influences upon river hydrology, such as water extraction and dam building, in our simulations. When these are taken into account, CSL could be expected to drop even further.

Our results suggest that the relatively warm period of the early-to-mid-Holocene is not a suitable analogue for the CSL under future global warming conditions. Although in both periods the simulated mean annual climate was warmer and wetter over mid-latitude Europe compared to the preindustrial era (here 1300–1750 CE), there are some important differences. The most important difference lies in the characteristics of the radiative forcing. In the early-to-mid-Holocene, insolation was slightly reduced during winter and increased during summer due to the nature of the orbital forcing, resulting in cooler winters (−0.5 to −1 °C) and warmer summers (+1.0 to +2.5 °C) compared to conditions for 1300–1750 CE (Renssen et al., 2005). In the 21st Century, on the other hand, radiative forcing is enhanced throughout the year due to elevated atmospheric greenhouse gas levels, producing a warmer annual mean climate than in the early-to-mid-Holocene (about 3 °C warmer over the Volga catchment in 2050–2100 CE). In the future climate, evaporation levels are therefore generally higher in the study area than in the early-to-mid-Holocene. As a result, the over-sea evaporative loss overwhelms the input from rivers and over-sea precipitation under the 21st Century anthropogenic scenario, while this is not the case for the 8–5.5 ka period (Fig. 4). Consequently, both our model and geological data show that the warm early Holocene climate is associated with relatively high CSL values (−21 mbsl), whereas the CSL is likely to experience a dramatic drop to below −30 mbsl in the warm climate of the late 21st Century.

3.4. Discussion of uncertainties and limitations

Our model experiment includes several uncertainties and limitations: (1) most importantly, our climate model has a low spatial resolution and simplified physics compared to GCMs. This makes it impossible to correctly simulate the spatial and temporal details of precipitation and evaporation in the study area, which could potentially be important for the calculation of CSL (e.g., Meshcherskaya et al., 1994; Meshcherskaya and Golod, 2003); (2) the impact of anthropogenic water extraction or other artificial constructions is not accounted for; if these are taken into account CSL could be expected to drop even further; (3) the low resolution forced us to statistically downscale our results to the 0.5° by 0.5° grid of STREAM, and to calibrate P_{CS} and E_{CS} based on modern estimates of over-sea precipitation and evaporation, in combination with simulated values over the Mediterranean Sea; (4) The 0.5° grid of STREAM allows only for delineating large river basins. Regrettably, the impact of these approximations cannot be estimated in this study, as this would require experiments with a more comprehensive, high-resolution model that is not yet available.

These limitations inevitably introduce uncertainties in our CSL calculations. It is clear that the CSL is extremely sensitive to small changes (i.e. in the order of 0.1 cm/yr) in the contributing climatic factors, i.e. over-sea precipitation and evaporation, and river runoff, which in turn depends on the hydrological balance over the catchments of the contributing rivers. Unfortunately, it is not feasible in this study to assess the accuracy of the contributing factors estimated by our model. This implies that our results should be treated with caution, and that they should be confirmed by independent model studies. The major advantage of our model setup is, however, that we are able to perform millennial-scale simulations of CSL changes, which is not yet feasible with GCMs.

Despite its limitations the model experiment: (1) is able to reproduce the main contribution to the inflow to the Caspian Sea accurately; (2) our simulated CSL variations are consistent with available data for various time-scales (Rodionov, 1994; Rychagov, 1997); (3) the results are generally consistent with recent model studies performed with state-of-the-art climate models (Elguindi and Giorgi, 2006a,b, 2007); (4) earlier studies have shown that our climate model captures the long-term Holocene climate trend well (Renssen et al., 2005), and the STREAM model could be well calibrated using the climate model data and is capable of simulating large-scale river runoff trends at various time-scales (Aerts et al., 2006; Ward et al., 2007).

4. Conclusions

We have simulated CSL variations for the period 8 ka to 2100 CE in a coupled climate-hydrological-sea level model. Our model results suggest the following:

The reconstructed long-term decrease in CSL can be attributed to a response to the orbitally-forced long-term reduction in summer insolation. Our model suggests a decline of 5 m from −21.5 mbsl between 8 and 5.5 ka to about −26.5 mbsl in the 18–19th Centuries CE. The magnitude of this downward Holocene CSL trend is in agreement with geological evidence for sea-level high stands. Long-lasting low stands with a CSL as low as −30 mbsl, as reported in the literature, are not reproduced by our model.

Internal variability in the climate system can produce sea level variations of about 4 m at centennial time-scales and about 2 m at decadal time-scales. Relatively small variations in precipitation and evaporation are amplified due to a high sensitivity of the CSL to subtle fluctuations in over-sea P–E and inflow from contributing rivers. The CSL fluctuations are consistent with historical and measured variations in CSL. The short-term variability increases after 6 ka.

The A1b anthropogenic emission scenario causes a drop in Caspian Sea level of about 4.5 m in the 21st Century mainly due to relatively strong over-sea evaporation. This drop is of the same order of magnitude as the orbitally-forced millennial-scale downward CSL trend simulated for the last 8000 years.

The relatively warm climate of the early-to-middle Holocene is not a suitable analogue for the end of the 21st Century, as over-sea evaporation becomes the dominant factor in the future, whilst during the period 8–5.5 ka over-sea evaporative loss was largely compensated by input from rivers and over-sea precipitation.

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