Climate change impact on the Mediterranean Sea circulation: a regional modelling approach

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1. Introduction

The global ocean thermohaline circulation (THC) is known to carry a large amount of the heat exchanged between the equator and the poles. Because of this crucial climate role (Broecker, 1991), the Atlantic THC has been largely studied. Indeed, its possible weakening could have an important impact on the European climate as mentioned in past-climate and future-climate studies (Ganopolski, 1998; Ganopolski and Rahmstorf, 2001; Clark et al., 2002; Vellinga and Wood, 2002; Weaver and Hillaire-Marcel, 2004). Moreover, the possibility of multi-equilibrium state (firstly shown by Stommel, 1961) and observations suggesting recent THC changes (Curry et al., 2003) have motivated a large number of studies about the stability of the THC in climate change scenario. A large range of climate model has been used (Stocker and Schmittner, 1997; Rahmstorf and Ganopolski, 1999; Wood et al., 1999) and a large range of THC answer has been obtained (IPCC, 2001). For example, Latif et al. (2000) find no THC change in their simulation whereas others (Manabe and Stouffer, 1999) obtain a 80% weakening.

In the Mediterranean Sea, many authors have also described a THC driven by heat and water losses at the sea surface (Wüst, 1961). This buoyancy flux leads to an anti-estuarine circulation with fresh and warm water getting into the Mediterranean Sea across the Gibraltar Strait at the surface and salty and cold water getting out in a deeper layer. With such a THC, at least one deep water formation area is needed as a key process for driving the vertical circulation. Actually, in the Mediterranean Sea, three main areas of intermediate or deep water mass formation have been reported: the Gulf of Lions also called MEDOC area (MEDOC Group, 1970; Schott et al., 1996) where the Western Mediterranean Deep Water (WMDW) is formed, the Levantine basin (eastern part of the Mediterranean Sea) where the Levantine Intermediate Water (LIW, Lascaratos et al., 1993) and the Levantine Deep Water (LDW, Ozsoy et al., 1993) are formed, and the Adriatic Sea (Artegiani et al., 1997) which is the main source of the Eastern Mediterranean Deep Water. Past-climate and present-climate studies have proved that the Mediterranean THC (MTHC) is very variable and even unstable. The sapropels formation is an evidence that this THC was shutdown or at least very weak in the past (Béthoux, 1993). More recently, the so-called Eastern Mediterranean Transient (EMT) has proved that the source of the EMDW can switch from the Adriatic Sea to the Aegean Sea during some years (Roether et al., 1996; Klein et al., 1999). This recent change seems to have many impacts on water mass structures and properties (Klein et al., 1999; Lascaratos et al., 1999; Manca et al., 2003) and on the biogeochemistry (Klein et al., 2003) in the Mediterranean Sea.

Following IPCC-A2 scenario, the climate over the Mediterranean basin may become warmer and drier during the 21st century (IPCC, 2001). These two effects could counteract each other in the Mediterranean Sea deep convection because the impact on the density of a warmer and saltier surface water is unknown. As for the Atlantic ocean, the weakening of the MTHC could have an impact on the Mediterranean sea surface temperature (SST), on the Mediterranean biogeochemistry and on the climate of the surrounding areas as well. Consequently, the possible evolution of the MTHC in a climate change scenario can be considered as a relevant question and we will try to assess it in this study.

Modelling the MTHC could be either easier or more difficult than modelling the global THC. On one hand, the MTHC time scale is smaller with a value around 70 years (Pickard and Emery, 1994) to be compared to 1000 years for the global one. Thus the impact of a MTHC modification can be visible more rapidly in the Mediterranean water mass characteristics than for the Atlantic. Besides, many authors (e.g. Béthoux et al., 1990; Rohling and Bryden, 1992; Fuda et al., 2002) reported temperature and salinity trends in the Mediterranean deep water and more recently in the Mediterranean outflow water in the Atlantic close to the Gibraltar Strait (Potter and Lozier, 2004). Béthoux et al. (1998) allocate these trends to global warming.

Moreover at a 70-year time-scale, we can assume that the impact of a MTHC modification on the Atlantic Ocean can be neglected and that we can study the MTHC changes without taking into account the feedback of these changes on the Atlantic Ocean.

On the other hand, modelling the MTHC could also be more difficult because it requires a higher horizontal and vertical resolution than for the global circulation. Small-scale advective and convective processes, small-scale atmospheric forcings and narrow straits have to be resolved indeed for allowing a good representation of the various water masses formation, advection and mixing (see Wu and Haines, 1996; Wu and Haines, 1998; Castellari et al., 2000; Brankart and Pinardi, 2001 for high resolution Mediterranean Sea modelling studies).

We also assume that small-scale structures of the SST modification do not have a major influence on large-scale structures of the atmospheric change in a scenario where the large-scale structures of Mediterranean SST modification are imposed.

The different time and space scales mentioned above and the different assumptions relative to the weak feedback of the Mediterranean Sea on global climate change lead to design the following experiment for assessing the question of the evolution of the MTHC: a high resolution ocean circulation model limited to the Mediterranean Sea is forced by air-sea fluxes coming from a previously and independently run scenario performed with a high resolution atmosphere regional climate model. SST anomalies outside the Mediterranean Sea or Atlantic water properties anomalies come from a coupled scenario done with a low resolution Atmosphere-Ocean

General Circulation Model (AOGCM). A 140-year IPCC-A2 climate change scenario (1960-2099) has been performed with this design. An equivalent control run under present-climate conditions has been carried out for checking the model stability.
Until now, past-climate studies and sensitivity studies of the MTHC have been done (Myers et al., 1998; Myers and Haines, 2002 ; Matthiesen and Haines, 2003) but, to our knowledge, no realistic climate change scenario for the 21st century with all forcings has been tested yet for the Mediterranean Sea.
The Mediterranean Sea model, its forcings and the simulations are presented in section 2. A present climate validation for the Mediterranean Sea model is shown in section 3. The climate change results focusing on the MTHC evolution are studied in section 4 and the results are discussed in section 5.

2. Models and simulations

A hierarchy of three different models is used to allow a dynamical downscaling of a climate change scenario of the Mediterranean Sea. The main model is a high resolution ocean regional circulation model (ORCM) limited to the Mediterranean Sea. It is forced by air-sea fluxes computed with a high resolution atmosphere regional climate model (ARCM). The ARCM is in turn forced by data from an atmosphere-ocean general circulation model (AOGCM). The links between the three models are displayed in the figure 1 and the different simulations are described below and summarized in table 1.

2.1 Mediterranean Sea model

A Mediterranean Sea limited area version of the primitive equation numerical model Ocean PArallel (OPA, Madec et al., 1998) has been developed. This model, called OPAMED8, is based on the 8.1 version of OPA and consequently is very close to the one developed for the MERCATOR project (Drillet et al., 2001; Bahurel et al., 2002; Béranger et al., 2004a; Béranger et al., 2004b; Alhammoud et al., 2004).
The horizontal eddy diffusivity and viscosity coefficients are fixed to -1.2 10^{10} m^4/s for tracers (temperature, salinity) and dynamics (velocity) with the use of a biharmonic operator. A 1.5 turbulent closure scheme is used for the vertical eddy diffusivity (Blanke and Delecluse, 1993) and the vertical diffusion is enhanced to 1 m^2/s in case of unstable stratification. The density is a function of the potential temperature relative to the sea surface, the practical salinity and the pressure (Jackett and McDougall, 1995). The C grid in Arakawa's classification (Arakawa, 1972) is used for the discretization. The bathymetry is based on the ETOPO5'x5' data base (Smith and Sandwell, 1997). The rigid lid hypothesis is applied at the surface and a free-slip lateral boundary condition is used. The bottom friction is quadratic. A time step of 20 minutes is used.
The area and the coast line of the model are presented in figure 4a. The horizontal resolution of OPAMED8 is 1/8°x1/8°cos(ϕ) with ϕ latitude. This is equivalent to a range of 9 to 12 km from the north to the south of the model with square meshes. It has 43 vertical Z-levels with an inhomogeneous distribution (from ΔZ = 6 m at the

surface to Δ Z = 200 m at the bottom with 25 levels in the first 1000 m). The maximum depth is 4100 m in the Mediterranean Sea (Eastern Basin).
The initial conditions are provided by the MEDATLAS-II monthly climatology for the Mediterranean Sea (MEDAR/MEDATLAS Group, 2002) and by a seasonal climatology (Reynaud et al., 1998) for the Atlantic part of the model. We start our simulations in August when the vertical stratification is the most stable to avoid a possible strong and non-physical initial mixing. The OPAMED8 grid is tilted and stretched at the Gibraltar Strait to better follow the SW-NE axis of the real strait. Then the Gibraltar Strait is represented with a two grid-point wide strait.
In this study, OPAMED8 is driven by three types of forcings :
1. At the surface, interannual water, momentum and heat fluxes are computed by a previously run atmosphere simulation. The atmosphere model and the simulations are described in the section 2.3. Air-sea fluxes change every day and are constant over a 24-hour period. Consequently, the diurnal cycle is not resolved by OPAMED8. Water fluxes coming from the atmosphere model are transformed into salt fluxes by dilution in the upper model level with respect to the rigid lid hypothesis. The heat flux is adjusted to the ORCM SST by a surface relaxation towards the daily SST used by the ARCM (section 2.4).
2. A salt flux due to river runoff is added to complete the salt budget (see section 2.5).
3. A buffer zone simulates the Atlantic Ocean. In this area, temperature and salinity are relaxed seasonally in 3D (see section 2.6).

2.2 Mediterranean Sea simulations : control run and scenario

For each forcing, we will define, in the next sections, present-climate and future-climate conditions. The present-climate conditions are used for forcing a 140-year control run, called Mediterranean Control (MC) and the future-climate conditions are used for a 140-year scenario, called Mediterranean Scenario (MS). Firstly, the MC run allows an evaluation of the stability of the ORCM over a long period of time. Secondly, the difference between MS and MC gives an unbiased evaluation of the climate change impact in case of a drift in MC.
A 20-year spin-up has been performed under present climate conditions before launching the MC and MS simulations in order to obtain a quasi-equilibrium. For this spin-up, OPAMED8 is forced two times successively by the interannual fluxes of the 1960-1970 period.

2.3 Air-sea fluxes

Atmosphere Regional Climate Model

Regional climate simulations can be performed with high-resolution AGCM (Cubasch et al. 1995), nested regional climate models (Giorgi and Mearns 1999), or statistical downscaling (Wilby et al. 1998). In the present study, we use the first method, which offers advantages in providing globally consistent simulations. However, in this experiment, high-resolution was restricted to the Mediterranean basin (Déqué and Piedelievre 1995).

In Déqué and Piedelievre (1995) and Machenhauer et al. (1998) we have shown that a variable resolution climate model realistically reproduces seasonal and geographical variations of the main climatological parameters over Europe. In Déqué et al. (1998) a 2xCO2 simulation was performed with a variable resolution version of ARPEGE.1.

The simulations used here have been performed with a new version of the model (ARPEGE.3). The two model versions are different in many respects. A description of ARPEGE.1 can be found in Déqué and Piedelievre (1995). ARPEGE.3 has been used in Gibelin and Déqué (2003) and we just recall here its features. It uses a semi-lagrangian advection with a two time-level discretization. The spectral truncation is T106, with 31 vertical levels located mainly in the troposphere(exactly those of ERA15 reanalysis, Gibson et al, 1997). The time step is 30 min. The pole of stretching is at the centre of the Mediterranean basin (40°N, 12°E) and the stretching factor is 3. The grid has 120 pseudo-latitudes and 240 pseudo-longitudes (with a reduction near the pseudo-poles to maintain the isotropy of the grid). As a result, the maximum horizontal resolution is 0.5°, that is to say about 50 km over the Mediterranean Basin, and reaches a minimum of 4.5° in the Pacific.

The convection scheme is derived from the mass-flux scheme with moisture convergence closure described by Bougeault (1985). The Morcrette (1990) scheme is used to calculate the radiation, which includes the effect of 4 greenhouse gases (CO2, CH4, N2O and CFC) in addition to water vapour and ozone, and of 5 aerosol types (land, sea, urban, desert and sulphate) in addition to background aerosols. Indirect effects of sulphate aerosols are parameterized by an empirical function for the cloud drop effective radius (Hu et al., 2001). The cloud-precipitation-vertical diffusion scheme uses the statistical approach of Ricard and Royer (1993). The soil scheme consists of a 4-layer diffusion scheme for temperature and the ISBA soil vegetation scheme (Douville et al., 2000) for the hydrological cycle. Representation of orographic gravity wave drag has been improved, with respect to the scheme used in version 1 (see Déqué et al., 1994), by the addition of mountain blocking and the lift effect (Lott and Miller, 1997; Lott, 1999).

1960-2099 Climate Simulation

The experimental set-up is different from Déqué et al. (1998), but exactly the same as in Gibelin and Déqué (2003). In the present study, the variable resolution model is run with radiative forcing (greenhouse gases and aerosols) following IPCC-A2 scenario (IPCC 2001). The only difference with Gibelin and Déqué (2003) is that an IPCC-B2 scenario was used there. This Atmosphere Scenario will be called AS in the following.

The SST/sea-ice forcing is based on a two-tier approach. From 1960 through 2000, we use interannual monthly mean observed SST, the so-called RSST, reconstructed with in-situ and satellite data (Smith et al., 1996). This allows validation of the variable resolution model by comparing the simulation with observations. The validation of years 1961-1990 against observations has been performed in Gibelin and Déqué (2003). Moreover, the difference between A2 and B2 scenarios starts only in 2000. The climate simulation over Europe is reasonably close to the CRU climatology (New et al., 1999). The largest deficiency is a too rainy winter. More details about the precipitation field can be found in Frei et al. (2003). From 2001

through 2100, artificial monthly SSTs were created by adding anomalies from a coupled run to observed SSTs.
Lower boundary conditions are given by a coarser resolution coupled simulation, and the atmosphere is assumed to be in equilibrium with the oceans and sea ice. The consistency of large-scale circulation patterns between ARCM and AOGCM simulations has been verified in Gibelin and Déqué (2003). Indeed, the physical parameterizations, which calculate the surface fluxes are the same in the coupled and the forced experiment.
In the present study, we use the same method to construct artificial monthly SSTs for 2000-2099 as in Gibelin and Déqué (2003), except that the radiative forcing is IPCC-A2 instead of IPCC-B2. This method uses unbiased 30-year monthly SST anomalies extracted from an AOGCM scenario performed with a lower resolution version of ARPEGE.3 (Royer et al., 2002; Douville et al., 2002).
In all this paper, the different years correspond to the years of the SST. Nevertheless these simulations are climate simulations and each model year is not a representation of the observed meteorological flow during this year due to the chaotic behaviour of the atmosphere. This remark is valid even for the present climate period.
The whole period of AS, 1960-2099, provides the air-sea fluxes for the Mediterranean Sea scenario (MS simulation). We have extracted daily momentum, water and heat fluxes from the AS simulation for the 1960-2099 period at a 0.5° resolution over the Mediterranean Sea. The OASIS2.4 tool (Valcke et al., 2000) is used to interpolate the fields from atmosphere to oceanic model. The 1960-1980 period of the AS simulation is used for forcing the control simulation (MC). This 20-year long forcing is repeated 7 times for obtaining a simulation as long as MS.

2.4 Consistency between surface heat flux and SST

For MC and MS, a newtonian relaxation is applied for the SST towards the SST used for forcing the ARCM run. The relaxation coefficient is -40 $W/m^2/K$, as defined in Barnier et al. (1995). It is equivalent to an 8-day restoring time scale, similar to those used in previous studies (e.g. Korres et al., 2000, 5-day restoring term). This term ensures a consistency between surface heat fluxes coming from the ARCM and SST calculated by the ORCM.

2.5 River runoff fluxes

No salinity damping is used at the surface of OPAMED8 because river runoff fluxes are applied explicitly. For the river runoff fluxes in MC, a monthly mean climatology is computed from the RivDis database (Vörösmarty et al., 1996) for the main 33 rivers of the Mediterranean Sea catchment basin.
The Black Sea is not included in OPAMED8. Nevertheless, this sea can be considered as one of the major freshwater sources for the Mediterranean Sea. Actually it is a salt sink in the model formulation. As for the Gibraltar Strait, the exchanges between the Black Sea and the Aegean Sea consist in a two layer flow across the Sea of Marmara and the Dardanelles Strait. In this study, we assume that this two layer flow can be approximated by a freshwater flux diluting the salt content of the mouth grid point. Thus the Black Sea is considered as a river for the Aegean Sea. The monthly mean equivalent water flux towards the Aegean Sea is computed as the water budget over the Black Sea surface : Precipitation + Black Sea River

Runoff – Evaporation. This parameterization is based on the data collected by Stanev et al. (2000) and assumes that the sea level of the Black Sea does not change. A more recent study (Tsimplis et al., 2004) shows that the Black Sea level increases. Consequently, our water budget should become: Precipitation + Black Sea River Runoff – Evaporation – Sea Level Increase. No 21st century scenario is available yet for Black Sea level and so this term is neglected.

For the MS run, a multiplying factor is computed to modify the river runoff monthly mean fluxes, depending on the river catchment. The TRIP runoff-model with 1° resolution (Oki and Sud, 1998) is forced by the ARCM 1960-2099 water fluxes (AS run). The daily runoffs of the Mediterranean rivers are averaged on ten years and we select eight main series to synthesize the behaviours of the rivers, namely the Rhone, the Po, Italian other rivers, the Ebre and Africa together, the Nile, Turkey, Greece, and the Black Sea.

Then running thirty years averages are computed, and the P-E average added to the Black Sea runoff, as mentioned above. Finally, the multiplying factor applied to the climatological monthly runoffs to extrapolate the river runoffs for the MS scenario is the division of this thirty-year average by the thirty first years average that is 1960-1989. The factors are updated each decade and we get an evolution of the river runoffs for the scenario which is coherent with the computation of SST used for relaxation. Table 2 presents the evolution of the factors from the 2000-2009 decade to the 2090-2099 decade. The main decrease is for the Black Sea, and we can expect an impact on the Aegean Sea. Note that for the Nile, the factor remains almost unchanged.

2.6 Buffer zone

The last forcing of OPAMED8 is a buffer zone which simulates the Atlantic Ocean west of the Gibraltar Strait. For MC and before 2000 for MS, in this area, temperature and salinity are relaxed towards seasonal 3D Reynaud climatology (Reynaud et al., 1998) thanks to a newtonian term in the tracer equation equal to -$(X_{model}-X_{climatology})/\tau$. The restoring term is weak close to Gibraltar (τ = 100-day time scale at 7.5°W) and stronger while moving away from it (τ = 3 days at 11°W).

For the scenario (MS), 3D temperature and salinity seasonally mean anomalies are computed from the ocean part of the AOGCM scenario. These anomalies are added to the Reynaud climatology after the same filtering as for the SST and the river runoff. This filtering means that, in the computed anomalies, trend signal and seasonal change signal are kept but that interannual change signal is removed.

3. Present climate validation

The main goal of this part is to show that OPAMED8 is stable over a long run. We also validate here OPAMED8 versus observations through the MC run for fields useful for representing the MTHC, namely air-sea fluxes, surface and deep temperature and salinity characteristics, mixed layer depth spatial pattern and water volumes transported by the MTHC.

3.1 Air-sea fluxes

Validating the air-sea fluxes is a difficult task because of the lack of spatial and temporal high resolution observed data. Nevertheless, for the Mediterranean Sea, some direct measurements exist and strait transport measurements give us indirect informations about surface fluxes. For example, the long term heat and water balance between the Gibraltar Strait transport and the Mediterranean Sea surface flux (Béthoux, 1979; Bunker, 1982) allows us to obtain indirect observations of the basin integrated value of the surface fluxes. Consequently, for validation, we focus on the 2070-2099 time and spatial average of the MC run over the whole Mediterranean basin. Then, air-sea fluxes are given for the main water mass formation areas in order to underline the specificity of these areas and to prepare the comparison with the scenario. Table 3 summarizes the air-sea fluxes values.

For the 2070-2099 period of the MC run, the Mediterranean surface heat loss is equal to -6.2 W/m^2 taking into account the SST relaxation term which amounts to 31.4 W/m^2. The -6.2 W/m^2 value is in good agreement with those found in the literature. From direct measurement, Béthoux (1979) gives indeed a value of -7 $\pm$ 3 W/m^2 for the surface heat flux. This value of -7 W/m^2 is confirmed by Bunker et al. (1982) whereas they found only +5 W/m^2 for the Gibraltar heat transport. Later measurements suggest a range of 5.3-6.2 W/m^2 (MacDonald et al., 1994) for the Gibraltar heat transport. In an ocean model study and including the SST relaxation term, Wu and Haines (1998) give a value of -5.8 W/m^2 for the two terms (40-year average after reaching a steady state). In the same kind of study, Castellari et al. (2000) obtain -9.8 W/m^2 for the surface flux (experiment D) and Brankart and Pinardi (2001) a value of -4 W/m^2. In the latter study, they forced their model with the 1945-1993 observed COADS heat flux (da Silva et al., 1995) with a damping term equal to -25 W/m^2/K. Disregarding SST relaxation, the COADS 49-year averaged heat flux is equal to +10 W/m^{-2}. In our case, the ARCM raw heat flux is equal to –37.6 W/m^{-2}. Our atmosphere model overestimates the observed value but it is, at least, negative and so simulates a heat loss as expected. Two other atmospheric datasets have been recently used by Josey (2003): the Southampton Oceanography Centre (SOC) flux climatology (Josey et al., 1999) with an integrated heat flux of +6 W/m^{-2} and the National Center for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) atmospheric model 1949-2002 reanalysis (Kistler et al., 2001) which gives a value of +2 W/m^{-2}. Consequently, the atmospheric datasets based on observed values, atmospheric reanalyses or climate models seem to be unable to produce accurate air-sea fluxes for the Mediterranean Sea. The SST relaxation term is thus necessary for forcing an ocean model. With this term, OPAMED8 is able to produce a relevant integrated heat flux.

The other important buoyancy forcing is the water flux (Evaporation – Precipitation – River input) at the surface. For this flux, the spread of the observed values found in the literature is larger than for the heat flux: the larger water loss is given by Béthoux (1979) with a deficit of 0.95 m/year and the smaller by Garett (1996) with 0.52 m/year. OPAMED8 gives a water loss of 0.72 m/year which is in the observed range and very close to the value given in Gilman and Garett (1994): 0.71 $\pm$ 0.07 m/year. Thanks to the GIBEX experiment at Gibraltar, Bryden and Kinder (1991) obtain a value of the Gibraltar Strait water transport which is equivalent to a E-P-R loss between 0.56 and 0.66 m/year. With the UNESCO database used in our study, the river runoff fluxes plus the Black Sea contribution is equal to 0.18 m/year and the E-P term is equal to 0.90 m/year. This value seems to be overestimated with respect to the latest values given by Josey (2003) for atmospheric datasets. He obtained E-P = 0.74 m/year for the SOC climatology and 0.70 m/year with the NCEP/NCAR reanalysis. However, in a modelling study, Wu and Haines (1998) obtained a value of -0.67 m/year with a surface salinity relaxation towards observed data. This value is very close to our E-P-R flux.

The stability of the surface fluxes as well as their interannual variability can be seen in figure 2 for the heat flux in W/m^2 and figure 3 for the water flux in mm/day for the 1960-2099 period of MC (black curve).

Table 3 summarizes the heat and water surface fluxes for the Mediterranean Sea and for various sub-basins where deep water formation may occur (Gulf of Lions, Levantine Basin, Adriatic Sea, Aegean Sea). Computing the buoyancy flux (see Marshall and Schott, 1999, equation (6), for the formula) allows us to understand the respective role of the water and heat terms in forcing the THC in each sub-basin. For the whole Mediterranean Sea, the buoyancy flux is negative in our notation. This means that water masses become denser due to air-sea fluxes over the Mediterranean Sea, that is to say that they lose buoyancy. Note that some authors express the buoyancy flux as a density flux in multiplying it by $-g/\rho$ with g the acceleration due to gravity and ρ the density (Josey, 2003).

The water forcing seems to be more important than the heat forcing with 70% of the total buoyancy flux. This average behaviour masks the seasonal and the interannual variability where the heat term dominates as mentioned by Josey (2003) for example. Thus, in winter, the heat term represents more than 80% of the total with a value of -38.7 10^{-9} $m^2 s^{-3}$. Besides, each sub-basin has different specificities. For example, the THC induced in the Gulf of Lions seems to be mainly due to the heat forcing. Indeed the yearly mean buoyancy flux due to the heat term is equal to -10.8 10^{-9} $m^2 s^{-3}$ which amounts to 2/3 of the total buoyancy flux. The situation is completely different in the Levantine Basin with 90% of the total buoyancy flux due to the water term. The Adriatic Sea and the Aegean Sea show a surface water gain due to the Po and Black Sea runoffs. In these sub-basins, the heat loss (negative buoyancy flux) dominates the water gain (positive buoyancy flux) in terms of buoyancy flux and it is responsible for the density increase and the sinking of the surface waters. Comparing the evolution of the water and heat term of the buoyancy flux under climate change conditions is one of the key-point in assessing the question of the MTHC future as will be shown in section 4.1.

3.2 Sea surface characteristics

The temperature and salinity sea surface characteristics are presented in table 4 averaged over the 2070-2099 period of MC, for the whole Mediterranean Sea and for various sub-basins. For the whole basin, we observed a cooling bias of -0.8°C with respect to the observed SST used for the relaxation for the same period: 19.5°C (Smith et al., 1996). The drift leading to this bias occurred during the OPAMED8 spin-up and then MC is very stable in SST from 1960 to 2099. This initial drift is due to the too strong Mediterranean net heat loss (-37.6 W/m^2 in average) simulated by ARPEGE without the SST relaxation. For the deep water convective areas, the bias is maximum in the Adriatic Sea with -1°C and minimum in the Gulf of Lions with -0.6°C.

The sea surface salinity (SSS) shows a very good agreement with the MedAtlas-II database (MEDAR/MEDATLAS-II group, 2002) with a bias of only 0.02 psu for the whole basin (38.18 psu for the 2070-2099 period of MC versus 38.16 for the data). This result validates the ARPEGE E-P-R flux of 0.72 m/year. The SSS bias is very weak in all sub-basins except for the Adriatic Sea where the bias is 0.67 psu. Because of SST and SSS biases, results concerning the Adriatic Sea have to be carefully taken into account. Nevertheless the general good agreement with the climatology has to be underlined, especially for the salinity, because this is the first Mediterranean Sea model forced by explicit river runoff fluxes without any relaxation towards SSS observed values (see Wu and Haines, 1998; Castellari et al., 2000; Brankart and Pinardi, 2001, e.g.).

The map of the SSS of OPAMED8 in winter (averaged over the last 30 years of MC) is shown in figure 4b and can be compared with the MedAtlas-II data in figure 4a. The areas of maximum SSS, namely the north of the Algero-Provençal basin, the Adriatic Sea and the north of the Levantine basin are the signature of the deep water formation areas. Comparing these two maps, the good agreement of the spatial patterns can be seen apart from the Adriatic Sea where the Po river freshwater input seems not to be large enough to decrease the SSS as in the climatology.

3.3 Heat and salt content

For studying the MTHC, not only the T-S surface characteristics have to be validated but also the heat and salt content of the model. We express these contents for various sub-basins by 3D temperature and salinity averaged from the surface to the bottom taking into account the volume of each grid box. The values are summarized in table 4 under S3D and T3D names. Comparing with the MedAtlas-II database, we show that our model is generally too cold (0.5°C in average over the whole basin) following the surface behaviour and that the 3D salinity is well reproduced with a bias equal to -0.01 psu. For the heat content, the most biased basins are the Aegean basin (-0.9°C) and the Adriatic sea (-0.8°C), the two shallowest basins. For the salt content, the maximum bias occurs in the Gulf of Lions with a fresh bias of -0.1 psu. This bias is probably due to a too weak LIW inflow at an intermediate depth in this area. This failing was already mentioned in other modelling studies (Wu and Haines, 1996; Castellari et al., 2000). It is interesting to note that the Adriatic sea does not show a large salt content bias despite its large SSS bias.

As for the surface characteristics, the major part of these bias appears during the 20-year OPAMED8 spin-up and figures 5 and 6 prove that the heat and salt contents of the model are remarkably stable over the 140-year control simulation for the whole Mediterranean Sea. The same behaviour is observed for all individual sub-basins (figures not shown).

3.4 THC characteristics

Methods

Validating the THC is difficult because of the lack of high resolution and long-term observed data at depth. However, the Mediterranean Sea has been observed for a long time and many authors have reported the geography of the different winter convective areas and the maximal depth that the convection can reach. Some of them tried to compute the water volume transported by the THC. In this section, we will validate these three approaches with OPAMED8.

Figure 7a represents the winter (JFM) mixed layer depth (MLD) averaged over the 2070-2099 period of the MC run. The MLD computation is based on a vertical eddy diffusivity criterion (limit = 5 cm^2/s). The shaded areas identify the areas of winter convection which are those mentioned in the literature, namely, the Gulf of Lions (MEDOC Group, 1970; Schott et al., 1996), the south of the Adriatic Sea (Artegiani et al., 1997), the Levantine basin (Lascaratos et al., 1993) and the south of the Aegean Sea (Roether et al., 1996). The winter convection is a process known to be very variable from one year to another (Marshall and Schott, 1999). So the MLD averaged value can mask different reality. Although it is beyond the scope of this study, the interannual variability of the maximum depth reached by the monthly mean MLD for each convective area is reported in figure 8a, 8b, 8c and 8d (black line).

Furthermore, we want to quantify the strength of the deep circulation for the water mass formation basins. For given boxes around these areas, we compute the horizontal water flux convergence/divergence across the box boundaries in Sverdrup (1 Sv = 10^6 m^3/s) for each model vertical level. This yields a 2D curve of the water transport convergence as a function of depth. At each level of a box, we identify inflow water (convergence, positive term) or outflow water (divergence, negative term). The vertically integrated value of the transport for each positive or negative part of the curve can be considered as an index of the MTHC strength. This method underestimates the actual value of each water mass transport because an inflow of a water mass can cancel out an outflow of another water mass at the same depth. However, we choose it for its simplicity and because it allows to quantify in Sverdrup the deep circulation changes between the control and the scenario. Moreover, the deep circulation is mainly driven by density contrasts and thus deep circulation changes will be understood as THC changes. Three sub-basins are considered: the Aegean Sea and Levantine Basin together limited at 21.7°E (figure 9a), the Adriatic Sea limited by the Otranto Strait (40°N - figure 9b) and the Gulf of Lions limited at 9.5°E and 40°N (figure 9c). For the Adriatic Sea, we have also defined a larger box limited at the north of the Ionian Sea (37.1°N – figure 9d) in order to check the sinking of the ADW after the Otranto Strait creating the EMDW. Table 5 summarizes the integrated values of each positive or negative part of the curves of figures 9a, 9b, 9c and 9d. These values can be identified as the inflow or the outflow of a specific water mass.

Levantine Intermediate Water

The Rhodes Gyre, area of the Levantine Intermediate Water formation, is well seen in figure 7a. The maximum winter MLD in figure 8a shows a large interannual variability. As mentioned by Ozsoy et al. (1993), the Levantine Deep Water is sometimes formed in the same area as the LIW. However, most of the time, the water mass sinking stops at an intermediate depth as expected.
Figure 9a shows a negative part at an intermediate depth (maximum at 400 m) which is mainly the signature of the eastward transport of the Levantine Intermediate Water. The integrated value of this transport is equal to 1.51 Sv. So, the LIW transport produced by OPAMED8 is in good agreement with previous studies. Indeed, Tziperman and Speer (1994) give a maximum LIW formation rate of 1.5 Sv from a surface climatology data study; Lascaratos et al. (1993) a value of 1.0 Sv with a mixed layer model and Castellari et al. (2000) a value of 1.5 Sv in a modelling study after modification of the surface heat flux formulae (DS1 experiment). With a 100-year simulation, Myers and Haines (2000) obtained a LIW formation rate of 1.2 Sv and a total eastward transport of 0.8 Sv across a vertical section at 25.5°E.

Adriatic Deep Water

In OPAMED8, the south of the Adriatic Sea is a well mixed area in winter as seen in figure 7a and the Adriatic Deep Water is created. Figure 8b shows a large interannual variability for the MLD maximum with a mean at about 800 m and some years for which the convection reaches the bottom. This behaviour corresponds to the observed one (Artegiani et al., 1997).
The flux of the ADW across the Otranto strait (negative part of the figure 9b) is equal to 0.47 Sv in agreement with Roether et al. (1994, 0.3 ± 0.1 Sv), Wu and Haines (1998, 0.44 Sv) and Castellari et al. (2000, 0.3 Sv).

Eastern Mediterranean Deep Water

The Eastern Mediterranean Deep Water is formed when the ADW overflows the sill at the Otranto Strait (Klein et al., 1999; Stratford and Haines, 2000).
In figure 9d, the negative bottom part of the curve represents the EMDW after the overflow. This figure combined with figure 9b proves that the ADW sinks between 40°N and 37.1°N to become the so-called EMDW. In our model, the EMDW transport maximum is situated between 1200 and 2300 m with non-negligible values until more than 3200 m. Castellari et al. (2000) obtain a 1700 m maximum sinking for the EMDW after the Otranto sill. The total transport is equal to 0.77 Sv higher than the ADW at Otranto because entrainment and mixing with LIW occurs during the overflow. However, this value of 0.77 Sv is underestimated with respect to the 1.5 Sv given by Tziperman and Speer (1994). This underestimation is mainly due to the computation method as mentioned above.
In figure 9a, the positive bottom part of the curve is the flow of the EMDW getting into the Levantine basin under the LIW current. We estimate it at 0.51 Sv. This proves the ability of OPAMED8 to ventilate the EMDW from its formation area to the rest of the basin. Myers ad Haines (2000) obtain a value of 0.1 Sv of the westward transport across a vertical section at 25.5°E associating it to the EMDW transport.

Even if the Aegean Sea is identified as a convective area in winter in our model simulation (see figure 7a and figure 8d), no EMT-like circulation is found in MC (Roether et al., 1996; Klein et al., 1999). Waters formed in the Aegean Sea in winter are not dense enough to overflow into the Levantine Basin and reach the bottom.

Western Mediterranean Deep Water

In figure 7a, the Gulf of Lions is shown as the major site of open-ocean deep water mass formation for the Mediterranean Sea as mentioned by Marshall and Schott (1999). For this area, figure 8a shows that the convection reaches a depth greater than 2000 m most of the time. Only some rare years do not show convection deeper than 500 m. Marshall and Schott (1999) precise that the two situations have been reported in the observations. In a modelling study, Castellari et al. (2000) obtain variable MLD from 400 m to the bottom of the basin as we do. However our convection in this area seems to be stronger than in their study because they obtain deep convection only every 3 years.

Figure 9c shows the LIW inflow into the Gulf of Lions area at an intermediate depth (Madec, 1990). From figure 9c, we evaluate it at 0.15 Sv but a large part of the LIW inflow is actually compensated by a WMDW outflow at the same depth but closer to the Spanish coast. Thus a large part of the LIW inflow is missing in the 0.15 Sv value. The bottom outflow formed by the newly formed WMDW is equal to 0.44 Sv. The difference between these two values is compensated by the surface currents: the eastern surface current from the Tyrrhenian Sea (+0.87 Sv, Liguro-Provençal Current) supplies more water in this area than the southern surface outflow (-0.58 Sv) takes away from it. When we compare with other studies, the WMDW transport value seems to be underestimated as for the LIW inflow. Indeed, Tziperman and Speer (1994) give a value of 1.0 Sv for the WMDW formation rate. From a modelling study for which the sea surface salinity climatology used for relaxation has been modified, Castellari et al (2000) obtain three different values with various parameterizations of the surface heat fluxes: 1.6 Sv (DS experiment), 1.1 Sv (DS2 experiment) and 0.2 Sv (DS1 experiment).

In conclusion, we have proved that OPAMED8 is able to produce a realistic as well as a stable THC for the Mediterranean Sea: this is true for heat and salt content, winter mixing areas geography, maximum mixed layer depth and deep circulation strength. These validating results allow some confidence in studying the possible evolution of the MTHC under IPCC-A2 scenario conditions.

4. Results of the climate change scenario

4.1 Air-sea fluxes

Table 3 summarizes the difference between MC and MS for the surface fluxes for the whole Mediterranean Sea and for the water mass formation basins. In addition to the heat, water and buoyancy fluxes validated in the previous sections, this table contains the yearly and winter averaged value of the wind stress norm (called tau in the following) and of the positive part of wind stress curl (called curl+ in the following). Indeed these terms are important in preconditioning the deep water formation (Madec et al., 1990; Marshall and Schott, 1999): Ekman pumping and isopycnes doming under cyclonic curl and mixing of the first ocean layers during a convective event. Changes in these two terms can clarify some situations not explained by the buoyancy flux.

For the whole Mediterranean Sea, the surface heat loss decreases from 6.2 W/m^2 to 1.8 W/m^2 and the water loss (or salt gain) increases from 0.72 m/year to 0.94 m/year (these values are averaged over the last 30 years of the runs). These results were expected because the Mediterranean climate is known to generally become dryer and warmer in an IPCC-A2 scenario (IPCC, 2001; Gibelin and Déqué, 2003). The time evolution of the heat and water fluxes of MC and MS are compared in the figure 2 and 3. The impact of these changes on the global buoyancy flux is quite negligible (see table 3). The buoyancy loss decreases by about 3% indeed and we are not able to answer the question of the evolution of the MTHC from this global approach. Looking at the tau and curl+ values, averaged over the whole basin, is not very relevant. However, we note that tau and curl+ decrease in winter by 25% and 15% respectively.

Focusing on different sub-basins (Gulf of Lions, Levantine Basin, Adriatic Sea, Aegean Sea), table 3 shows that the yearly mean heat loss decreases everywhere except for the Aegean Sea where it increases. Table 3 shows also that the surface water loss increases everywhere. This last feature is particularly interesting for the Adriatic Sea and the Aegean Sea where the buoyancy flux due to the water flux is positive in MC and negative in MS. This means that evaporation is lower than precipitation plus river runoff in MC and higher in MS. This result is mainly due to the decrease of the river runoff in our A2 scenario, the Po river for the Adriatic Sea and the Black Sea for the Aegean Sea.

The effect of the two parts of the buoyancy flux are opposed except for the Aegean Sea where the buoyancy loss increases by 93%. Remember that a negative buoyancy flux leads to a surface density increase. Thus a buoyancy loss increase between MC and MS means a stronger density increase. For the other basins, the opposition leads to a decrease of the buoyancy loss for the Gulf of Lions (-21%), a small and probably non significant increase for the Levantine Basin (+7%) and a larger increase for the Adriatic Sea (+24%). Moreover, table 3 shows that the forcing due to tau and curl+ decreases in winter in every sub-basins except in the Aegean Sea.

This surface fluxes analysis leads to assess the reaction of the winter deep convection in these different sub-basins. For the Gulf of Lions, the forcings tend to a weakening of the winter convection and so a weakening of the western part of the MTHC. On the contrary, the winter convection should be enhanced in the Aegean Sea with saltier and colder surface waters and a more windy weather particularly in winter. The situations of the Levantine Basin and of the Adriatic Sea have to be checked more carefully because of antagonistic forcings.

4.2 Warming and salting

Table 4 summarizes the comparison between MC and MS in terms of SST, SSS, heat content (T3D) and salt content (S3D). For the SST, the mean warming is equal to +2.5°C and it is quite homogeneous over the entire basin. This is due to the SST relaxation applied towards data coming from the AOGCM run. This 8-day time-scale relaxation allows the formation of local patterns in the SST response but no

large-scale difference between the model SST and the low resolution SST. A high resolution Mediterranean Sea-Atmosphere coupled model is needed for further improvement concerning the high resolution SST feedback to the atmosphere.
For the SSS, the model answer is more spatially heterogeneous. The basin-scale average is an increase of 0.33 psu with areas of lower increase (Gulf of Lions: +0.28 psu) and areas of larger increase (Adriatic Sea: +0.61 psu and Aegean Sea: +0.70 psu). The river runoff flux decrease is the main cause of the Aegean Sea and Adriatic Sea behaviour. These SSS and SST lead to lower surface densities in MS than in MC for every sub-basins. This shows that the surface density evolution is not only driven by the buoyancy flux which increases in some sub-basins as seen before. At least, the advection of lighter waters from other sub-basins plays an important role. The Aegean Sea (-0.17 $kg.m^{-3}$) and the Adriatic Sea (-0.23 $kg.m^{-3}$) are the two sub-basins where the density does not decrease a lot. The Gulf of Lions (-0.41 $kg.m^{-3}$) and the Levantine Basin (-0.49 $kg.m^{-3}$) show more important surface density decrease. For each sub-basins, SSS, SST and surface density changes have the same behaviour in winter (values not shown) as in average over the year. This leads to conclude that deep water formation should decrease with respect to the surface density changes especially in the Gulf of Lions and in the Levantine Basin. However, the Adriatic Sea and the Aegean Sea might keep a strong local vertical circulation. This spatial discrimination is mainly due to the SSS changes which are mainly driven by river runoff water supply changes. Indeed, the changes of the Evaporation-Precipitation term is spatially homogeneous (figure not shown).
The heat and salt content changes represent the ability of OPAMED8 to transfer the surface anomalies towards the deeper layers by the different vertical and horizontal physical processes (vertical mixing, subduction, diffusion, advection). For the whole Mediterranean Sea and for each sub-basin, table 4 shows areas where this transfer is more efficient (Adriatic Sea) or less efficient (Levantine Basin) than the average. This allows the following hypothesis: in the scenario, the deeper waters of the Levantine Basin are weakly ventilated whereas the deeper layers of the Adriatic Sea seem strongly ventilated. This could be the signature of the weakening of either the EMDW formation by cascading or the EMDW advection. For explaining in details these ventilation differences, the contribution of surface fluxes, mixing and advection have to be further analysed for each sub-basins but this is not the topic of this study.

4.3 Thermohaline circulation weakening

The comparison between figure 7a and figure 7b shows that the winter averaged MLD has decreased in all convective areas at the end of the MS run. The winter averaged MLD is only weakly modified for the Adriatic Sea and for the Aegean Sea as expected from the analysis of the air-sea fluxes and of the surface characteristics. The Levantine Basin does not show any higher than 200 m MLD in MS but this maximum was already weak in MC. The most impressive point is the answer of the Gulf of Lions area. The winter averaged MLD do not exceed 300 m in the Gulf of Lions in the MS simulation whereas it was deeper than 1500 m in MC. We note that no other deep water formation area appears in MS.
This first qualitative aspect is confirmed by the analysis of figures 8a, 8b, 8c and 8d which show the yearly time series of the monthly mean maximum MLD for MC and

MS for the different sub-basins. The Adriatic and Aegean Sea winter convection is weakly changed. Comparing the 2070-2099 average of this variable for the two simulations, we estimate the winter convection weakening at -9% for the Adriatic Sea and -15% for the Aegean Sea. The role of the large SSS increase in these areas seems to be confirmed. For the Levantine basin, the weakening is evaluated at -57% by the same method. Nevertheless, figure 8a proves that this strong weakening is due to a decrease in LDW formation frequency. Indeed, the LIW formation remains yearly even if it occurs with a shallower maximum depth. For the Levantine Basin, the new configuration without LDW formation is stable since 2020 in MS. The Gulf of Lions presents a similar behaviour with a rapid weakening of the frequency of the very deep convection years (> 1500 m) and a stabilized situation after 2020. However, at the end of the scenario, convection deeper than 1000 m is still possible. Over the last 30 years, the weakening is evaluated at -76%.

The weakening of the winter deep water formation intensity should have an impact on the MTHC. Figure 9a, 9b, 9c, 9d and table 5 show the comparison between MC and MS in terms of convergence and divergence of the horizontal circulation which is a simple and good index of the THC as explained above. The major conclusion is that the MTHC becomes a shallow circulation instead of a deep circulation. Indeed, intermediate waters own a smaller but still important volume transport whereas the deep and bottom waters become almost motionless. The LIW volume transport decreases by 23% just after its formation area (outflow from the Levantine Basin centred around 400 or 500 m in figure 9a), by 37% when it gets into the north of the Ionian Sea (inflow towards the Adriatic Sea centred at 400 m in figure 9d) and by 33% when it gets into the Gulf of Lions area (inflow centred at 750 m). Despite this weakening by one third of the intermediate THC, the Gibraltar Strait exchanges do not evolve a lot with only a 6% decrease of the volume transport from 1.10 Sv to 1.03 Sv (figure not shown).

For the deep or bottom water masses, the situation is more critical with an averaged decrease in volume transport by more than 75%. Even if the ADW is always formed with the same rate (about 0.5 Sv for a 30-year average, see figure 9b), the formation of the EMDW decreases by 38% (0.77 Sv in MC versus 0.48 Sv in MS at the north of the Ionian Sea, see figure 9d). This is the signature of a density decrease of the water created by mixing of the newly formed ADW with the northwards LIW inflow. This mixed water, called EMDW in present climate, is less able to cascade after the Otranto Sill in the scenario. Figure 9d shows, for example, that the EMDW maximum transport is situated at 2200 m in MC and only at about 1000 m in MS. This is more evident at 21.7°E where the EMDW transport has decreased by 75% before entering into the Levantine Basin (see figure 9a). Moreover, the maximum transport is at 1100 m in the scenario instead of 2500 m in the control run. Finally, we conclude to a large weakening of the Eastern Mediterranean THC in our scenario. No sign of any long-term EMT-like circulation is found in our scenario even if the Aegean Sea has been identified as a potential deep water formation site through the air-sea fluxes analysis. Note that the small amount of deep water transport towards the west in the scenario at 21.7°E between 2000 and 3000 m is too weak (0.06 Sv, see figure 9a) to be significant. However, figure 8d shows that a convection deeper in MS than in MC can occur during some very rare winters such as 2011, 2038 and

2058. A study of these specific years has to be performed before a definitive conclusion about EMT-like circulation and climate change in the Mediterranean Sea.
In the Western Basin (Gulf of Lions area, figure 9c), the situation is similar. On average over the last 30 years of the simulations, the WMDW outflow towards the south is reduced by 82% from 0.44 Sv to 0.08 Sv. Moreover, in MC, the water divergence, representing the domination of the WMDW transport at a given depth, is maximum between 2000 and 2500 m whereas it shows its maximum transport before 1500 m in MS. The weakening of the deep Western MTHC is so very strong and it is accompanied by a smaller weakening of the surface and intermediate component of the Western MTHC : the LIW transport decreases by 33% and the Liguro-Provençal current by 38%.
A 1960-2099 time-series study should be done in another paper to clarify the time lag-correlation between the different water masses transport decreases. Indeed, the following questions remain opened: Does the LIW formation decrease lead or lag the WMDW or EMDW formation weakening ? Does the WMDW weakening lead or lag the Liguro-Provençal current weakening ? These questions seem crucial for better understanding the answer of the Mediterranean water masses system to the climate change. These issues could also allow to assess and understand the current climate time-variability of the Mediterranean water mass formation.
In conclusion, the impact of an IPCC-A2 climate change scenario on the Mediterranean thermohaline circulation seems to be a shallowing of the main deep water masses and a water masses transport weakening of about 30% for the intermediate circulation and about 75% for the deep and bottom circulation.

5. Concluding remarks

Conclusion

We performed a realistic scenario of what could be the 21st century of the Mediterranean Sea under IPCC-A2 scenario hypothesis. The various forcings (air-sea surface fluxes, river runoff fluxes and Atlantic-Mediterranean exchanges) have been taken into account using previously run AOGCM and ARCM. For the whole Mediterranean Sea, the heat loss by the surface decreases from 6.2 to 1.8 W.m^{-2} and the water loss (or salt gain) increases from 0.72 to 0.94 m/year. The wind stress norm and the positive part of wind stress curl decrease in the studied sub-basins except in the Aegean Sea.
A spatially homogeneous SST increase (+2.5°C for the end of the 21st century) is obtained whereas an heterogeneous SSS increase is produced by the model (from +0.28 psu in the Gulf of Lions to +0.70 psu in the Aegean Sea). The SSS anomalies pattern is mainly driven by the river runoff decrease and especially the Po and the Black Sea behaviour.
These changes lead to a decrease in the surface density and thus a weakening of the Mediterranean thermohaline circulation. This weakening is evaluated to about 75% for the bottom and deep circulation (WMDW, EMDW) with a very strong signal in the western basin and to 30% for the intermediate circulation (LIW). In the Adriatic Sea, the formation of the ADW remains quite stable. No EMT-like event is observed

in our scenario though the climate change fluxes over the Aegean Sea should favour deep water formation in this basin.

The salinity and temperature surface anomalies are transmitted into the deeper layers. This transmission is more efficient in the shallow basins which keep a vertical THC (Adriatic Sea, Aegean Sea) than in the deeper basins (Levantine Basin, Gulf of Lions area). On average over the whole Mediterranean Sea, the heat content increase is 1°C and the salt content increase is 0.18 psu. Even though it is not the topic of this study, we can expect that these T-S modifications should change the Mediterranean Outflow Water (MOW) characteristics flowing into the Atlantic Ocean across the Gibraltar Strait. Warming and salting of the MOW have been reported from hydrographic data for the last decades (Potter and Lozier, 2004) and might be already a Mediterranean Sea climate change signature.

In this study, we have used a control run (or present-climate run) of same length as the scenario for evaluating a possible drift of our model. The stability of this run as well as the weak bias shown by our model allow some confidence in the scenario results.

Uncertainties

However, in the current study, we did not explore the uncertainties linked to climate change scenarios. Indeed, as mentioned in the European PRUDENCE project (Prediction of Regional scenarios and Uncertainties for Defining EuropeaN Climate change risks and Effects, Christensen et al. 2002), many sources of uncertainties are related to the projection of regional climate changes. To improve the confidence level of our results, other IPCC scenarios and other atmospheric forcings have to be tested.

Moreover, many authors proved the high sensibility of the Atlantic THC to ocean model parametrizations, resolution and complexity and also to THC initial state. Thus, we think that ensemble simulations are needed for assessing in details the possible evolution of the Mediterranean THC for the 21st century as in the case of the global THC.

River runoff impact

In the scenario, the spatial heterogeneity of the SSS modifications is mainly driven by the runoff decrease of the rivers of the southern Europe. This heterogeneity plays a major role in influencing the spatial pattern of the MTHC weakening. Nevertheless, the river runoff is known to be very difficult to reproduce by a GCM (Douville et al., 2002) even with a 50 km resolution (Hagemann and Jacob, 2004). Moreover, in an ensemble of IPCC-A2 scenarios, Hagemann and Jacob (2004) have proved that rivers show a broad range of answers to the climate change in different regional climate models. This confirms again the need for ensemble simulations for which we could use river runoff fluxes anomalies coming from various regional climate models. For this purpose, the integrating role of our Black Sea/Aegean Sea exchanges parametrization should not be underestimated. Indeed, this parametrization integrates the uncertainties related to the model water flux over the Black Sea plus the precipitation over the Black Sea catchment basin.

Impact of the SST relaxation

In a forced ocean model with SST relaxation, the model SST is mainly driven by the damping SST which has to be known *a priori*. In our scenario, we use a low resolution AOGCM scenario for creating the damping SST anomalies. The use of this constraint supposes that the Mediterranean SST large-scale answer to the climate change is comparable in OPAMED8 and in the low resolution ocean model of the AOGCM. This hypothesis is probably true to the first order but we think that the regional pattern of the MTHC weakening could have a negative feedback to the atmosphere through the SST. This feedback will be taken into account in a future work with a high resolution Mediterranean-atmosphere coupled model already used in present climate studies (Sevault et al., 2002; Somot et al., 2002).

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simulations	AS	MC	MS
models	ARPEGE	OPAMED8	OPAMED8
years	1960-2099	1960-2099	1960-2099
resolution (Med. Sea)	50 km	10 km	10 km
control / scenario	Scenario from 2000	control	Scenario from 2000
GHG and aerosols	obs then IPCC-A2	-	-
Air-sea fluxes	-	1960-1980 AS	1960-2099 AS
SST	obs then obs + AOGCM ano	obs	obs + AOGCM ano
runoff	-	obs	obs + AS ano
buffer zone	-	obs	obs + AOGCM ano

Tab. 1: Characteristics of the different simulations used in this study, AS for Atmosphere Scenario, MC for Mediterranean Control run and MS for Mediterranean Scenario (we use obs for observations and ano for anomalies).

Rivers	2000	2010	2020	2030	2040	2050	2060	2070	2080	2090
Rhone	0.98	0.95	0.97	0.93	0.92	0.90	0.89	0.86	0.81	0.80
Po	0.92	0.86	0.84	0.79	0.82	0.82	0.79	0.79	0.79	0.81
Italy others	1.05	1.02	0.94	0.86	0.87	0.90	0.91	0.91	0.88	0.88
Ebre and Africa	0.93	0.75	0.80	0.75	0.74	0.72	0.64	0.64	0.46	0.44
Nile	1.08	1.09	1.10	1.04	1.06	0.94	1.00	0.98	1.03	0.93
Turkey	0.79	0.81	0.77	0.78	0.75	0.70	0.61	0.47	0.44	0.43
Greece	1.04	0.96	0.88	0.78	0.79	0.83	0.79	0.75	0.66	0.60
Black Sea	0.86	0.75	0.70	0.61	0.55	0.58	0.57	0.50	0.36	0.29

Tab. 2: Factors applied to the climatological monthly runoffs during one decade, for the eight chosen series.

Basin	***Mediterranean Sea***				
Flux	***Qtot*** ***W/m²***	***E-P+R*** ***m/year***	***Buoy (heat+water)*** ***10^{-9} m².s⁻³***	***Tau (JFM)*** ***10^{-2} N/m²***	***Curl+ (JFM)*** ***10^{-7} N/m³***
MC	- 6.2	0.72	-9.8 (-3.0-6.8)	3.21 (5.87)	1.40 (2.19)
MS	- 1.8	0.94	-9.5 (-0.9-8.6)	2.67 (4.38)	1.41 (1.87)
MS-MC	+ 4.4	+0.22	+0.3 (+2.1-1.8)	-0.54 (-1.49)	+ 0.01 (-0.32)
MS-MC	+71%	+31%	+3%	-17% (-25%)	+1% (-15%)

Basin	***Gulf of Lions***				
Flux	***Qtot*** ***W/m^2***	***E-P-R*** ***m/year***	***Buoy (heat+water)*** ***10^{-9} $m^2.s^{-3}$***	***Tau (JFM)*** ***10^{-2} N/m^2***	***Curl+ (JFM)*** ***10^{-7} N/m^3***
MC	-22.43	1.56	-16.2 (-10.8-5.4)	5.19 (9.20)	2.37 (3.49)
MS	-12.42	1.98	-12.7 (-6.0-6.7)	4.10 (6.49)	2.22 (3.23)
MS-MC	+10.01	0.42	+3.5 (+4.8-1.3)	-1.09 (-2.71)	-0.15 (-0.26)
MS-MC	+45%	+27%	+22%	-21% (-29%)	-6% (-7%)

Basin	***Levantine Basin***				
Flux	***Qtot*** ***W/m^2***	***E-P+R*** ***m/year***	***Buoy (heat+water)*** ***10^{-9} $m^2.s^{-3}$***	***Tau (JFM)*** ***10^{-2} N/m^2***	***Curl+ (JFM)*** ***10^{-7} N/m^3***
MC	-2.26	2.81	-10.6 (-1.1-9.5)	3.37 (5.29)	1.33 (2.29)
MS	-0.29	3.29	-11.3 (-0.1-11.2)	2.77 (4.15)	1.31 (2.01)
MS-MC	+1.97	+0.48	-0.7 (+1.0-1.7)	-0.60 (-1.14)	-0.02 (-0.28)
MS-MC	+87%	+17%	-7%	-18% (-22%)	-2% (-12%)

Basin	***Adriatic Sea***				
Flux	***Qtot*** ***W/m^2***	***E-P-R*** ***m/year***	***Buoy (heat+water)*** ***10^{-9} $m^2.s^{-3}$***	***Tau (JFM)*** ***10^{-2} N/m^2***	***Curl+ (JFM)*** ***10^{-7} N/m^3***
MC	-16.86	- 0.23	-7.9 (-8.1+0.2)	1.66 (2.46)	4.05 (4.81)
MS	-15.64	0.54	-9.8 (-7.5-2.3)	1.46 (1.73)	2.39 (3.39)
MS-MC	+1.22	0.77	-1.9 (+0.6-2.5)	-0.20 (-0.73)	-1.66 (-1.42)
MS-MC	+ 7%	+335%	-24%	-12% (-30%)	-41% (-30%)

Basin	***Aegean Sea***				
Flux	***Qtot*** ***W/m^2***	***E-P-R*** ***m/year***	***Buoy (heat+water)*** ***10^{-9} $m^2.s^{-3}$***	***Tau (JFM)*** ***10^{-2} N/m^2***	***Curl+ (JFM)*** ***10^{-7} N/m^3***
MC	-21.38	- 1.33	-9.9 (-10.3+0.4)	4.03 (4.27)	2.94 (3.03)
MS	-28.14	1.12	-19.1 (-13.5-5.6)	4.76 (5.46)	2.90 (3.04)
MS-MC	-6.76	+2.45	-9.2 (-3.2-6.0)	0.73 (1.19)	-0.04 (-0.01)
MS-MC	-32%	+184%	+93%	+18% (+28%)	-1% (0%)

Tab. 3: Sub-basins time and space averaged air-sea fluxes over the 2070-2099 period for the control run (MC), the scenario (MS) and the difference between them. For the buoyancy flux, the terms due to heat flux and to water flux are indicated in brackets. For tau and curl+, the winter averaged values are in brackets.

Basin	***Mediterranean Sea***			
°C or psu	***SST***	***T3D***	***SSS***	***S3D***
MC	18.7	13.2	38.18	38.61
MS	21.2	14.2	38.51	38.79
MS-MC	+2.5	+1.0	+0.33	+0.18

Basin	***Gulf of Lions***			
°C or psu	***SST***	***T3D***	***SSS***	***S3D***
MC	16.8	12.4	37.97	38.31
MS	19.3	13.5	38.25	38.48
MS-MC	+2.5	+1.1	+0.28	+0.17

Basin	***Levantine Basin***			
°C or psu	***SST***	***T3D***	***SSS***	***S3D***
MC	20.1	13.7	39.04	38.84
MS	22.8	14.5	39.37	38.97
MS-MC	+2.7	+0.8	+0.33	+0.13

Basin	***Adriatic Sea***			
°C or psu	***SST***	***T3D***	***SSS***	***S3D***
MC	17.0	13.0	38.43	38.60
MS	19.7	15.1	39.04	39.08
MS-MC	+2.7	+2.1	+0.61	+0.48

Basin	***Aegean Sea***			
°C or psu	***SST***	***T3D***	***SSS***	***S3D***
MC	17.9	13.9	38.45	38.85
MS	20.6	15.6	39.15	39.22
MS-MC	+2.7	+1.7	+0.70	+0.37

Tab. 4: Sub-basins time and space averaged SST (in °C), heat content (in °C), SSS (in psu), salt content (in psu) over the 2070-2099 period for the control run (MC), the scenario (MS) and the difference between them.

Basin	***Gulf of Lions***			
Current	***Eastern surface***	***Southern surface***	***intermediate***	***deep***
MC	+ 0.87 Sv	- 0.58 Sv	+ 0.15 Sv	- 0.44 Sv
MS	+ 0.54 Sv	- 0.57 Sv	+ 0.10 Sv	- 0.08 Sv (*)
MS-MC	- 38%	- 2%	- 33%	- 82%

Basin	***Adriatic Sea (Otranto Strait - 40°N)***	
Current	***surface***	***deep***
MC	+ 0.47 Sv	- 0.47 Sv
MS	+ 0.48 Sv	- 0.48 Sv
MS-MC	+ 2%	+ 2%

Basin	***Adriatic Sea (Ionian Sea - 37.1°N)***	
Current	***surface***	***deep***
MC	+ 0.78 Sv	- 0.77 Sv (*)
MS	+ 0.49 Sv	- 0.48 Sv (*)
MS-MC	- 37%	- 38%

Basin	***Levantine Basin and Aegean Sea (21.7°E)***		
Current	***surface***	***intermediate***	***deep***
MC	+1.00 Sv	- 1.51 Sv	+ 0.51 Sv
MS	+ 1.06 Sv	- 1.16 Sv	+ 0.13 Sv (*)
MS-MC	+ 6%	- 23%	- 75%

Tab. 5: Sub-basins time averaged horizontal water mass transport (in Sv) over the 2070-2099 period for the control run (MC), the scenario (MS) and the difference between them. The values correspond to integrated values over each converging or diverging part of the figures 9a, 9b, 9c and 9d and are called surface, intermediate and deep with respect to these curves. (*): the sum of the terms does not amount to zero because small parts of the curve are not mentioned in this table.