Multiple equilibria and overturning variability of the Aegean-Adriatic Seas

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Abstract

The Eastern Mediterranean Transient (EMT)–a transition and amplification of the Eastern Mediterranean Sea deep water source from the Adriatic Sea to the Aegean Sea–was observed in the mid-90’ and stimulated intense research. Here we demonstrate, using an oceanic general circulation model, that the meridional overturning circulation of the Eastern Mediterranean has multiple equilibria states under present-day-like conditions, and that the water exchange between the Aegean and the Adriatic Seas can drastically affect these states. More specifically, we found two stable states and a hysteresis behaviour of deep water formation in the Adriatic Sea when changing the atmospheric (restoring) temperature over the Aegean Sea. In addition, the overturning circulation in both seas exhibits large decadal variability of the deep water formation. The Aegean-Adriatic relationship can be summarized as follows: warm and saline water of the Aegean can either flow in the sub-

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surface to the Adriatic, switching “on” deep water formation in the Adriatic by increasing its salinity, or the Aegean water can feed the deeper layer of the Ionian and Levantine basins, turning “off” the deep water formation in the Adriatic. The “off” steady state resembles some aspects of the EMT in which the Adriatic source of deep water was weakened when the Aegean source became active.

**Keywords:** Mediterranean Sea, Meridional overturning circulation, Multiple equilibria, Adriatic Sea, Aegean Sea, Eastern Mediterranean Transient.

1. Introduction

The Eastern Mediterranean Transient (EMT)—a transition of the major deep water formation site in the Eastern Mediterranean from the Adriatic Sea to the Aegean Sea—is an intriguing change in Mediterranean circulation that was observed in the instrumental era. For almost a century, the Adriatic Sea was the Eastern Mediterranean (EM) major source of deep water (Nielsen, 1912; Wüst, 1961). The Aegean Sea was considered a potential, but sporadic, source of dense water (Roether and Schlitzer, 1991). Surprisingly, a new and much stronger source of deep water was found in the Aegean during a hydrographic survey in 1995 (Roether et al., 1996), an event that was termed the EMT. The EMT attracted the attention of the scientific community that proposed various explanations for this transition (Lascaratos et al., 1999; Malanotte-Rizzoli et al., 1999; Theocharis et al., 2002; Josey, 2003). Understanding the causes and nature of abrupt changes such as the EMT is important since deep water formation processes allow the exchange
of physical and biochemical properties (e.g., oxygen and nutrients) between
the surface and the deep layers. It possibly also affects the regional climate
by changing the stability and the structure of the water column over a wide
area.

The Adriatic and Aegean deep water formation sources are different in
their nature. When deep water formed in the Adriatic, before the EMT,
the average formation rate was $\sim0.3$ Sv ($1$ Sv $= 10^6$ m$^3$s$^{-1}$). The Adriatic
water sank to the bottom of the Ionian (Roether and Schlitzer, 1991), and
from there spread southward and eastward (Wüst, 1961; Malanotte-Rizzoli
et al., 1997), filling the abyssal EM basins. The water of the new Aegean
source, formed during the EMT, was denser (warmer but saltier) than the
Adriatic deep water and formed at a mean rate of $\sim1$ Sv (about three times
the Adriatic deep water formation rate). The new Aegean deep water out-
flowed through the Crete Straits toward the Ionian and Levantine basins, and
by the year 1995 replaced almost $20\%$ of the old EM deep water (Roether
et al., 1996; Malanotte-Rizzoli et al., 1997). It uplifted the nutrient level into
the euphotic zone and enhanced EM biological productivity (Klein et al.,
1999; ?). Apparently, the Aegean deep water outflow started around 1987
(Theocharis et al., 1999) and peaked in 1993 (Roether et al., 2007). The
Aegean source weakened after the EMT and the Adriatic source recovered
(Manca et al., 2003; Velaoras and Lascaratos, 2005; CIESM, 2012) while ex-
hibiting large variability in its deep water properties (Cardin et al., 2011,
2014).

Several explanations for the EMT have been proposed and there is an
ongoing discussion whether the EMT was a single phenomenon that was
caused by the unique conditions in the EM during 1987-1992 (Theocharis et al., 1999; Zervakis et al., 2000; Krokos et al., 2014), or whether it is a recurrent phenomenon associated with the natural variability of the EM circulation (Borzelli et al., 2009; Gačić et al., 2010; Theocharis et al., 2014; Velaoras et al., 2014).

For instance, it was suggested that the EMT was stimulated by the intense heat loss during the particularly cold winters of 1991-1993 (Lascaratos et al., 1999; Wu et al., 2000; Stratford and Haines, 2002; Josey, 2003; Beuvier et al., 2010), by intensification of the prevailing wind-stress (Samuel et al., 1999), or by altered intermediate water circulation (Wu and Haines, 1996; Malanotte-Rizzioli et al., 1999; ?). Malanotte-Rizzioli et al. (1999) observed that during the EMT, an anticyclonic circulation between the Levantine and the Ionian basins was blocking the salty Levantine Intermediate Water (LIW) inflow into the Adriatic Sea. Other studies focused on the increase of salinity (and hence the density) in the Aegean Sea, enabling the EMT. It was associated with a decrease in precipitation over the Aegean (Theocharis et al., 1999); with a reduction in the inflow from the Black Sea (Zervakis et al., 2000); and with the long-term salinity increase in the EM due to the damming of the Nile river (Skliris et al., 2007). Few studies suggested that both an increase in Aegean salinity and extreme cold weather conditions are essential to cause an EMT-like event (Lascaratos et al., 1999; Krokos et al., 2014).

Studies linking the EMT to natural variability of the EM circulation, include, e.g., the Adriatic-Ionian Bimodal Oscillating System (BiOS) mechanism (Gačić et al., 2010). According to this mechanism, a cyclonic circulation in the north Ionian causes an advection of saline LIW into the Adriatic, en-
hancing deep water formation there. In contrast, anticyclonic circulation in the north Ionian leads to advection of fresher Modified Atlantic Water (MAW) into the Adriatic, weakening the Adriatic deep water formation. This mechanism is associated with a feedback between the LIW, MAW, and the stretching of the water column due to changes in water density. Another mechanism, suggested by Theocharis et al. (2014), is the “pumping mechanism” in which the outflow of dense water through deep layers of either the Otranto Strait or Crete Straits is balanced by the inflow of MAW in the upper layers. The relatively fresh MAW stabilizes the water column of the sea into which it flows, halting (or significantly weakening) its deep water formation. At the same time, the other source is enhanced by receiving LIW inflow instead of MAW, eventually leading to a switch from one source of deep water to the other.

Stommel (1961) raised the possibility of multiple equilibria states of the overturning circulation in the Mediterranean Sea, similar to the ones he predicted using a simple two-box model in the Atlantic Ocean. Recently, Ashkenazy et al. (2012) showed, using a three-box model representing the Adriatic, Aegean, and Ionian Seas, that the meridional overturning circulation of the EM may indeed have a few steady states. Ashkenazy et al. (2012) also associated the EMT with a transition between these states.

Previous studied examined the EM nonlinear response to linear changes in external (atmospheric) forcing using Ocean General Circulation Models (OGCMs) with different levels of complexity. For example, Myers and Haines (2002) forced a coarse resolution OGCM with realistic atmospheric conditions and found that the Mediterranean zonal overturning circulation collapses
when increasing (+25%) or decreasing (-20%) the evaporation over the entire Mediterranean. Meijer and Dijkstra (2009) used an OGCM forced by paleo-atmospheric conditions to study the ventilation of the EM sediments in periods equivalent to sapropel episodes (Rohling, 1994; ?; ?). They found an abrupt, transient halt of the Mediterranean deep zonal overturning circulation under a decreased meridional temperature gradient. Pisacane et al. (2006) used mixed boundary conditions (i.e., specified freshwater flux and temperature restoring), and found that the EM exhibits large self-sustained variability on a decadal time scale. In their model, the EM undergoes internal oscillations of weakening and strengthening of the zonal overturning circulation cell as a result of alternation between the Adriatic and the Aegean sources. Pisacane et al. (2006) related the EMT to an advection-convection feedback mechanism, similar to the later work of Theocharis et al. (2014), rather than to an atmospheric cause.

Here we performed a set of experiments using an OGCM to demonstrate that the overturning circulation in the Adriatic Sea has multiple steady (or quasi steady) states under the same atmospheric conditions. Moreover, the overturning circulation in the Adriatic exhibits a hysteresis behaviour with respect to the atmospheric (restoring) temperature over the Aegean Sea, i.e., it responds to distant atmospheric variations. Essentially, we found in the Adriatic two steady states of overturning circulation with and without an active source of deep water. The two steady states exist under minor atmospheric temperature anomalies (relative to present-day conditions) over the Aegean. Different quasi steady states were found also in the Aegean under the same atmospheric conditions, however not as distinct as in the Adri-
Figure 1: The Mediterranean Sea bathymetry used in the OGCM simulations. The red line indicates the transect shown in Fig. 2. The red squares indicate the position of the profiles shown in Fig. 8.

We also found that the meridional overturning circulation in both the Adriatic and Aegean Seas experience decadal variability that resembles the deep decoupling oscillations of Winton and Sarachik (1993). Given the above we thus conjecture that EMT-like event may occur spontaneously without a major change in atmospheric or other forcing.

The paper is organized as follows: in section 2 we describe the main elements of our model and the experimental steps. In section 3 we present our results and we conclude in section 4.
2. Model and experimental design

2.1. Model setup

We use the Massachusetts Institute of Technology OGCM (the MITgcm, Marshall et al., 1997a,b) to perform the simulations. The MITgcm supports a wide range of parameterizations and boundary layer options. We use the finite-volume, z-coordinate, hydrostatic, free surface, partial cell options of MITgcm. The simulated domain consists of the entire Mediterranean Sea area, covering the Strait of Gibraltar from the west (Fig. 1).

Model bathymetry is based on ETOPO2 (Smith and Sandwell, 1997). The bathymetry at the Strait of Gibraltar was widen and deepen since the resolution of the model is not fine enough to reproduce the observed water flux through the strait. The water column is resolved by 22 vertical levels centered at: 10, 30, 50, 75, 100, 125, 150, 200, 250, 300, 400, 500, 600, 800, 1000, 1200, 1500, 2000, 2500, 3000, 3500 and 4000 meters. The horizontal resolution is $1/8^\circ \times 1/8^\circ$ on a spherical grid, which is equivalent to 14 km in the meridional direction and to 9-12 km in the zonal direction. Since the first internal Rossby radius of deformation is 12-15 km for most of the Mediterranean Sea (Robinson et al., 1992), our simulations can be considered as an eddy-permitting simulations. The time step for both the tracers and the momentum is 1200 seconds.

Ocean convection was simulated using the implicit vertical diffusion coefficient scheme of MITgcm, i.e., enhancement of the vertical diffusivity to 1 m$^2$s$^{-1}$ in regions where the water column becomes unstable. The background vertical eddy diffusivity was simulated through a Laplacian formulation with diffusivity coefficient ranging from $3 \times 10^{-4}$m$^2$s$^{-1}$ at the sur-
face to $1 \times 10^{-6} \text{m}^2\text{s}^{-1}$ at the bottom. The vertical viscosity coefficient is $1.5 \times 10^{-4} \text{m}^2\text{s}^{-1}$ over the entire water column. Horizontal salinity and temperature diffusive terms were modelled using the bi-harmonic formulation of MITgcm, with diffusion coefficients equal to $1.5 \times 10^{16} \text{m}^4\text{s}^{-1}$. Smagorinsky harmonic viscosity coefficient equal to 3 was set to formulate the horizontal viscosity according to Griffies and Hallberg (2000).

2.2. Model spin-up

The model was initialized with climatological 3D temperature and salinity from NEMO-MED reanalysis (obtained by Copernicus Marine Service Products, http://marine.copernicus.eu/). The wind-stress was constructed from the monthly climatology of ERA-Interim reanalysis (Dee et al., 2011). The Atlantic Ocean input was simulated by prescribing Levitus (1982) monthly climatology of temperature and salinity in the western boundary of the model, outside the Strait of Gibraltar. This was done using the restoring boundary condition package (RBCS) of MITgcm. All initial and forcing fields were spatially interpolated or extrapolated to fit the model resolution.

In the control run, for 1500 years, sea surface temperature (SST) and sea surface salinity (SSS) were relaxed toward the monthly averaged climatological SST and SSS of NEMO-MED reanalysis with a relaxation time of 8 and 6 days, respectively. Then, monthly salt/freshwater flux was diagnosed and the simulation continued with mixed surface boundary conditions (of prescribed freshwater flux and relaxation to NEMO-MED SST) for additional 500 years, to allow the development of natural variability in the simulations (Haney, 1971).

In Fig. 2 we show the temperature and salinity of the control run, of
NEMO-MED reanalysis, and the differences between them, along a transect in the EM (shown in Fig. 1). The low-salinity signature of the MAW appears in the western side of the transect while the deeper, saline signature of the LIW appears in the eastern side of the transect. The position of these two distinct water masses is not exactly as their position in the reanalysis, but their main properties are well represented. Despite the differences in temperature and salinity over the EM transect, water fluxes through the Sicily, Otranto and Crete Straits are found to be well within the observed range, reproducing the EM internal variability reasonably well.

The overturning circulation is often used to quantify deep water formation. Fig. 3 depicts the control run Meridional Overturning Circulations (MOCs) of the Adriatic (a) and the Aegean (b), indicating a deep Adriatic MOC cell and a shallow Aegean MOC cell. The Aegean annual mean outflow of water denser than 1029.2 kg m$^{-3}$ (the most dense water in the EM) in the control run is 0.01 Sv while the Adriatic outflow is 0.5 Sv. Therefore, consistent with observations, the major deep water formation source in the control run is the Adriatic Sea.

2.3. Experimental design

The experiments were inspired by studies that investigated the multiple equilibria of the Atlantic Ocean MOC as a function of freshwater forcing (Stommel, 1961; Manabe and Stouffer, 1988; Marotzke and Willebrand, 1991; Stouffer et al., 2006). Alike these studies, MOC hystereses curves were found when we continually increased or decreased a boundary condition parameter between extrema values, as a function of time. Unlike these studies, we choose the thermal forcing (restoring temperature) over the Aegean and
Figure 2: EM annual averaged temperature and salinity depth transects (along the red line in Fig. 1) of the control run (a and b) vs. NEMO-MED reanalysis (c and d) and the differences between the model and the reanalysis (e and f).
Adriatic Seas as the control parameter; see Ashkenazy et al. (2012). This was motivated by the observed severe winter cooling, of almost 3°C, over the Aegean Sea during the EMT (Josey, 2003). Hence, larger forcing bounds between restored temperatures were chosen to detect the temperature range in which two MOC states arise.

In the first (second) experiment, hereafter Exp1 (Exp2), we modified the Aegean (Adriatic) temperature boundary condition, i.e., the restoring temperature over the Aegean (Adriatic), but kept the rest of the forcing and boundary conditions as in the control run. Both experiments were conducted in several steps as detailed below.

- **Step 1:** We started from the control steady state and restored the Aegean (Adriatic) SST to temperature that are 5°C colder than the control run temperature, until reaching a steady state (after ~600 years). This state is termed “the cold steady state”.
• Step 2: We then increased the restoring temperature over the Aegean (Adriatic) by 10°C at a rate of 0.02°C/yr. This step thus lasted 500 years and reached a restoring temperature of 5°C above the control run restoring temperature. This step is termed “Gradual Warming” (GW).

• Step 3: Then, we kept the Aegean (Adriatic) restoring temperature 5°C warmer than the control run restoring temperature and continued the simulation till reaching a steady state (after ~600 years). This state is termed “the warm steady state”.

• Step 4: We started from the warm steady state and decreased the restoring temperature of the Aegean (Adriatic) at a rate of 0.02°C/yr for 500 years, until it reached again the restoring temperature that is 5°C colder than the control run. This step is termed “Gradual Cooling” (GC).

• Step 5: Finally, we kept the Aegean (Adriatic) restoring temperature 5°C colder than the control run restoring temperature and continued the simulation till reaching a steady state (typically after ~600 years).

The above steps aimed to construct hysteresis curves of the Adriatic and Aegean overturning circulation with respect to the restoring temperature. To explore possible existence of multiple stable (quasi) steady states, we ran in addition the model for selected restoring temperatures in Exp1, starting with initial conditions from the GW and GC simulations (steps 2 and 4 above). Specifically, we performed the following experiments:

1. We applied control forcing and boundary conditions for 1500 years, starting from the state achieved in the GW under zero temperature
anomaly during step 2. We termed this experiment “From Gradual Warming” (Exp1 FGW).

2. We applied control forcing and boundary conditions for 1500 years, initializing from the state achieved in the GC under zero temperature anomaly during step 4. We termed this experiment “From Gradual Cooling” (Exp1 FGC).

3. Same as 1 but with $-0.2^\circ$C temperature anomaly over the Aegean. We termed this experiment “From Gradual Warming -0.2” (Exp1 FGW.m0p2).

4. Same as 2 but with $-0.2^\circ$C temperature anomaly over the Aegean. We termed this experiment “From Gradual Cooling -0.2” (Exp1 FGC.m0p2).

Thus, in total we constructed two hysteretic curves (in Exp1 and in Exp2) and four steady state experiments (only from stages of Exp1), as summarized in Table 1:

<table>
<thead>
<tr>
<th>Simulation name</th>
<th>Simulation duration</th>
<th>Initial conditions</th>
<th>Restored T-anomaly over Aegean/Adriatic</th>
</tr>
</thead>
<tbody>
<tr>
<td>GW</td>
<td>500 years</td>
<td>cold steady state</td>
<td>from $-5^\circ$C to $5^\circ$C</td>
</tr>
<tr>
<td>GC</td>
<td>500 years</td>
<td>warm steady state</td>
<td>from $5^\circ$C to $-5^\circ$C</td>
</tr>
<tr>
<td>FGW</td>
<td>1500 years</td>
<td>year 250 of GW sim.</td>
<td>zero anomaly</td>
</tr>
<tr>
<td>FGC</td>
<td>1500 years</td>
<td>year 250 of GC sim.</td>
<td>zero anomaly</td>
</tr>
<tr>
<td>FGW.m0p2</td>
<td>1500 years</td>
<td>year 240 of GW sim.</td>
<td>$-0.2^\circ$C</td>
</tr>
<tr>
<td>FGC.m0p2</td>
<td>1500 years</td>
<td>year 260 of GC sim.</td>
<td>$-0.2^\circ$C</td>
</tr>
</tbody>
</table>

Table 1: Simulations summary
3. Results

To quantify the deep water formation in the GW and GC steps of Exp1 and Exp2, we calculated the outflow from Otranto and Crete Straits of water denser than 1029.1 kg m\(^{-3}\). The LIW density is \(\sim\)1029 kg m\(^{-3}\) (Lascaratos et al., 1993), and so denser water that outflow from the Adriatic and Aegean Seas is forming the deep water mass of the EM. The outflow from those seas is southward, hence more negative values represent more dense water outflow.

However, quantifying the deep water formation using somehow arbitrary density threshold provides only a partial view as the flux of deep water depends on this threshold. To complement the picture we also compared buoyancy gain or loss for both seas. We calculated Buoyancy Flux (BF) through the Otranto and Crete Straits as follows:

\[
BF = g \int_A \rho v dA
\]

where \(g\) is the gravitational acceleration, \(v(x, y, z, t)\) is the meridional, across-strait velocity, \(\rho(x, y, z, t)\) is water density, and \(A\) is the strait cross-section area. Thus, the BF can be calculated for each point along the axis perpendicular to the cross-section. The BF through a strait indicates how the water density varies with time, hence its potential to produce deep water. If the BF is positive then the sea ’gains’ mass by dense water flowing into it, while if the BF is negative, the sea produces deep water that outflows from it to possibly fill (if the water is dense enough) the deep layers of the Ionian and Levantine basins. As more negative it is, the sea produces additional mass as more or/and denser deep water.
3.1. Hysteresis

3.1.1. Exp1

In the first step of Exp1, a cold steady state is achieved by reducing the Aegean restoring temperature to be 5°C colder than in the control run. This steady state is characterized by an outflow of 0.07 Sv from the Adriatic (Fig. 4a, red line at −5°C) and 1.6 Sv from the Aegean (Fig. 4b, red line at −5°C). The Aegean surface cooling results in cold and hence dense water that sinks and outflows as deep water through Crete Straits. In this state,
the water entering the Adriatic is denser than that leaving it (positive BF in Fig. 4c at $-5^\circ$C), and there is almost no dense water outflow. The warm steady state achieved in step 3 is very different (Fig. 4, at $5^\circ$C). The Aegean surface water is too warm to sink to bottom layers so there is almost no outflow (Fig. 4b), while deep water is formed in the Adriatic as indicated by the outflow of 0.6 Sv (Fig. 4a). These results highlight the feedbacks between the Aegean and Adriatic Seas, where, e.g., changes in the restoring temperature over the Aegean drastically affect the Adriatic outflow.

Note that the GW and GC curves do not necessarily merge in the extreme negative values of their temperature forcing (Fig. 4b, at $-5^\circ$C and Fig. 5a, at $-5^\circ$C). This is since the red line at $-5^\circ$C represents the cold steady state (not explicitly shown, i.e., we start the GW simulation after reaching a cold steady state) while the blue line at $-5^\circ$C represents the state of the cooling phase at this point. At the end of the cooling step we again run the model to a cold steady state and the curves merged (not shown).

The Adriatic and Aegean dense water outflow and BF during steps 2 (red line) and 4 (blue line) are shown in Fig. 4 (a,c and b,d respectively). The most prominent result is that the Adriatic outflow exhibits hysteresis despite the fact that the boundary conditions were modified only over the Aegean. Such a hysteresis curve suggests that there are two Adriatic steady states under the same Aegean forcing. Furthermore, the largest difference in the outflow is found around zero forcing anomaly in the Aegean restoring temperature, i.e., under present-day conditions. The Aegean outflow also exhibits a hysteresis curve around zero forcing anomaly in response to changes in its forcing, however it is over a smaller range of temperature anomalies and it does not
Figure 5: Adriatic and Aegean annual mean outflow of water denser than 1029.1 kg m\(^{-3}\) (a and b, respectively) and BF (c and d, respectively) as a function of the temperature anomaly of the Adriatic restoring temperature in Exp2. Gradual warming (GW, red line) and gradual cooling (GC, blue line) steps are plotted.

3.1.2. Exp2

In the first step of Exp2 the Adriatic was forced by restoring temperature anomaly of \(-5^\circ\)C and run to a steady state. This steady state is characterized by an outflow of 1 Sv from the Adriatic (Fig. 5a, red line at \(-5^\circ\)C) and 0.75 Sv from the Aegean (Fig. 5b, red line at \(-5^\circ\)C). The BF from the Adriatic in that steady state is negative (Fig. 5c), indicating deep water formation. In contrast, the BF from the Aegean (Fig. 5d) is positive. Therefore, even
though there are 0.75 Sv of dense water outflow, the water inflow is denser as revealed by the BF values. This means that the inflow occurs in the deep layers of Crete Straits and the outflow in the shallow layers.

The warm steady state achieved in step 3 is very different (Fig. 5, blue line at 5°C); the Adriatic surface water is too warm to sink to bottom layers so the outflow is minimal, indicating that it is not a deep water formation source. However, the Aegean outflow of dense water is also generally small, with some outflow episodes (Fig. 5b, blue line at 5°C). The negative BF of the Aegean (Fig. 5d at 5°C) indicates that its water inflow is less dense than the outflow (active MOC, but weak). However, the water outflow from the Adriatic is not dense enough to sink to the bottom of the EM, and in this state none of the seas produce dense enough water that can ventilate the deep layers of the EM.

In Exp2, the Adriatic outflow exhibits a small amplitude hysteresis behaviour and the Aegean outflow exhibits somewhat larger amplitude hysteresis behaviour, e.g., in −2°C, than in Exp1. However, when running these transient states to stable steady states, they converge (not shown).

3.2. Steady states stability analysis

3.2.1. Zero forcing anomaly

To examine the stability of the two Adriatic MOC states shown in Fig. 4, we took the model transient states from Exp1 GW and Exp1 GC steps under zero forcing anomaly, and ran them for additional 1500 years under the control boundary conditions. The results are shown in Fig. 6 for the Adriatic and Aegean (a,c and b,d respectively). Both simulations kept their initial Adriatic outflow (BF) for approximately 600 years, until in Exp1 FGW the
Figure 6: Time series of Adriatic and Aegean annual mean outflow of water denser than 1029.1 kg m$^{-3}$ (a and b, respectively) and annual averaged BF (c and d, respectively), of the FGW and FGC experiments; i.e., under zero anomaly forcing (same boundary conditions as in the control run) and initial conditions are the zero anomaly states in Exp1 GW and GC.
outflow (BF) abruptly changed and merged with the one of Exp1 FGC simulation. Before that point, the natural variability in the FGC is much larger than in the FGW simulation. Around year 1200 of the FGW simulation there is another abrupt (but transient) change of a drastic weakening of the Adriatic outflow (BF) for almost a decade.

While there is a general agreement between the Adriatic dense water outflow time series and its corresponding BF time series, the Aegean dense water outflow time series (Fig. 6b) is very different from its BF time series (Fig. 6d). In the FGW simulation, the Aegean dense water outflow exhibits large variability (order of 0.7 Sv) and abrupt changes, similar to the ones in the Adriatic. However, the Aegean BF is similar in both FGW and FGC cases with mostly negative low values that indicate a shallow MOC as in the control run. We next review in details the abrupt changes in the FGW simulation in both seas.

To examine the interesting transitions that appear in Exp1 FGW simulation (Fig. 6), temperature, salinity, and density time series of the intermediate (200-500 m) and deep (500-1500 m) layers of the Aegean and Adriatic are plotted (Fig. 7). At first (years 0-600), the deep layers of the Aegean warm gradually as a result of heat diffusion from the surface and intermediate layers, leading to a weaker stratification with time. Note that the Exp1 GW step began from the cold steady state, and the slow warming of the surface did not penetrate to the deep layers immediately. The deep layers of the Aegean are also getting saltier, approaching the salinity values of the intermediate water (38.8 psu). However, the temperature effect on density is more dominant in this period (Fig. 7f). Around year 600, the Aegean destabilized and
Figure 7: Annual averaged temperature (a and b), salinity (c and d) and density (e and f) time series of the intermediate [200-500 m] (a, c and e) and deep [500-1500 m] (b, d and f) layers of the Aegean and Adriatic Seas through Exp1 FGW simulation.
a complete mixing occurs, leading to twice as large BF, indicating that dense water leaves the sea. As a result, after this abrupt transition, the bottom water of the Aegean is warmer and saltier due to the enhanced MOC.

Up to this transition point, the MOC of the Adriatic is not active (positive BF and no dense water outflow in Fig. 6a,c). Although pulses of warm water do occur (Fig. 7b), leading to some destabilization, the overall trend in the Adriatic bottom water during years 0-600 is of cooling and freshening. Here, the salinity is more dominant in the destabilization of the Adriatic water column and accordingly, the Adriatic MOC is initiated (Fig. 6). At and after the transition (around year 600), the Adriatic deep and intermediate water becomes abruptly saline (Fig. 7c,d), indicating an external source of salinity.

After the transition (that occurred around year 600), both the Aegean and Adriatic bottom water salinity is increased, leading to enhanced stratification. At year $\sim$850 the water column of the Aegean becomes stable and the MOC is terminated for a short time. For the same reason, the Adriatic MOC weakens (the BF is less negative). The ceasing of Aegean MOC leads to fresher intermediate and bottom water in both the Aegean and Adriatic. This freshening destabilizes the water column (between years 850-950), gradually recovering back the MOC. Then, diffusion of salt from the surface weakens again the stratification till year $\sim$1150, at which full mixing occurs. The process described above is repeated with a major difference in the Adriatic: its MOC is terminated at year $\sim$1200 (instead of the weakening that occurred around year 850). This relaxation oscillations like cycle is repeated several times and it seems to be triggered by the Aegean intermediate saline outflow. This process is similar to the relaxation oscillations mechanism of

To better understand the connection between the Aegean and Adriatic properties, we retrieved density profiles in the south Adriatic and south Aegean Seas (at the locations marked by the red squares in Fig. 1) over years 1170-1270 of Exp1 FGW simulation (during which the abrupt transient occurred in the Adriatic). It is evident from Fig. 8 that a freshening of intermediate south Aegean water occurred prior to and during the Adriatic transition. It seems that dense (saline) water associated with the LIW (200-700 m) is found only beneath 400 m starting from year 1190. This water is known for its role in deep water formation source water in the Adriatic (Lascaratos et al., 1999). High salinity peak is evident in the Aegean bottom water (Fig. 7d) before the Adriatic MOC transient to a passive state.

By tracking passive tracer released in the intermediate (200-500 m) layers of the Aegean Sea we can see that before the transition of year 1200 (Fig. 9a-c), the Aegean intermediate saline water was ‘feeding’ the Adriatic and enabled its deep MOC. In this case, the Adriatic outflow propagated along the coasts of the north Ionian. Differently, during the transition itself (Fig. 9d-f), the Aegean intermediate water does not flow into the Adriatic and instead outflowing from Crete Straits to fill the Levantine and Ionian basins, as been observed during the EMT.

3.2.2. −0.2°C forcing anomaly

Similar to the above numerical experiments, we initiate the model from the two branches of the hysteresis curve (i.e., from Exp1 GW and Exp1 GC steps) under −0.2°C forcing anomaly over the Aegean, and ran them for additional 1500 years. These are the FGW.m0p2 and FGC.m0p2 described in
Figure 8: Density profile time series in the Adriatic (a) and in the Aegean (b) over the years 1170-1270 of Exp1 FGW. Profiles locations are marked by red squares in Fig. 1.
Figure 9: Passive tracer concentration one (a and d), three (b and e) and five (c and f) years after it was released in the intermediate (200-500 m) layers of the Aegean at year 1180 (upper panels) and at year 1200 (lower panels) of Exp1 FGW simulation.
Figure 10: Time series of Adriatic and Aegean annual mean outflow of water denser than 1029.1 kg m$^{-3}$ (a and b, respectively) and annual averaged BF (c and d, respectively), of the FGW.m0p2 and FGC.m0p2 experiments; i.e., under $-0.2^\circ$C temperature anomaly forcing and initial conditions are the $-0.2^\circ$C anomaly states in Exp1 GW and GC.
Sec. 2. The outflow and BF time series of the Adriatic (Fig. 10a,c) indicate two separate stable steady states under the same forcing of $\sim 0.2^\circ$C temperature anomaly over the Aegean. One of the states is of positive BF, indicating a passive MOC in the Adriatic in the FGW.m0p2 simulation, while the second state is of negative BF, indicating an active MOC in the FGC.m0p2 simulation. Note that the FGC.m0p2 state is highly variable (order of 0.6 Sv). The Aegean BF’s (Fig. 10d) in both simulations are very similar and merge into a single steady state. However, the Aegean dense water outflow (Fig. 10b) time series are highly variable and exhibit different and statistically distinct steady states. The FGW.m0p2 simulation has smaller amplitude of outflow in which most of the time it is less than 0.5 Sv, while the FGC.m0p2 simulation exhibits values of up to 1 Sv (as was observed during the EMT) with a few episodes of outflow ceasing.

The last 10 years average of each steady state MOC (Fig. 11) show that indeed the Adriatic MOC is very different in both states while the Aegean MOC states are similar. However, a strengthening of the Aegean MOC is evident when the Adriatic MOC is shut down. Still, the outflow through the Aegean in the two states is different; see Fig. 10.

Similarly to the transient feature around year 1200 of Exp1 FGW experiment, passive tracer released in the intermediate (200-500 m) layers of the Aegean Sea in each of the steady states is spreading very differently. When the Adriatic MOC is active (Fig. 12a-c), Aegean intermediate water is entering the Adriatic and maintains its deep water formation active state. When the Adriatic MOC is passive (Fig. 12d-f), there is no inflow from the Aegean into the Adriatic. Since the Aegean MOC is somewhat more active at this
Figure 11: Ten years averaged MOC from the FGW.m0p2 and FGC.m0p2 experiments in the Adriatic (a and c, respectively) and Aegean (b and d, respectively).
Figure 12: Passive tracer concentration one (a and d), three (b and e) and five (c and f) years after it was released in the intermediate (200-500 m) layers of the Aegean at year 1500 (upper panels) of FGC.m0p2 simulation and at year 1500 (lower panels) of FGW.m0p2 simulation.
stable steady state, a spreading of its dense water to fill the abyssal layers of the Levantine is detected instead.

4. Conclusions

In summary, we investigated the hypothesis that multiple equilibria MOC states exist in the Adriatic-Aegean Seas, using a state-of-the-art OGCM. We only varied one atmospheric variable so far, the restoring temperature, and found multiple MOC states in the Adriatic and to a lesser extent in the Aegean. Obviously, variation of other atmospheric variables, such as the freshwater flux and wind stress, may also result in multiple MOC states, both in the Adriatic and in the Aegean. It is worth noting that sensitivity experiments (different rate of cooling/warming, finer vertical resolution) yielded similar results (not shown), confirming the robustness of the presented results.

The main result of this work is that under present-day-like atmospheric conditions, the EM overturning circulation in an OGCM has multiple equilibria. The two steady states are of an active Adriatic MOC (dense water outflowing from the Otranto Strait) and of a passive Adriatic MOC (the water enters the Adriatic are denser than those leaving it). The passive Adriatic MOC state is a result of an initialized colder Aegean (Exp1 GW) that is kept forced by minor temperature anomaly over the Aegean (Exp1 FGW.m0p2). To our knowledge, the passive Adriatic state has never been predicted before and may be relevant to abrupt changes in the EM. A possible multiple equilibria in the Aegean dense water formation is also implied but is not as distinct as in the Adriatic and need to be further pursued.
The second result of this work is that under present-day atmospheric conditions the EM overturning circulation exhibits decadal variability and abrupt transients, similar to the ones reported by Pisacane et al. (2006). More specifically, the Adriatic MOC can shutdown for a decade or so under the same annual atmospheric conditions (as in year 1200 of Exp1 FGW).

The third result is the relationship between the Aegean and Adriatic Seas. First, we demonstrate that changes over one sea induce variations in the other as well. Second, we explain how this link is established via horizontal advection. It appears that when the Aegean stops supplying saline intermediate water to the Adriatic, the Adriatic MOC becomes passive. This result is similar to the “pumping mechanism” suggested by Theocharis et al. (2014) with the exception that the Adriatic outflow does not reach the Aegean. We detect a mainly unidirectional influence between the two seas. Furthermore, when a transition from the quasi steady state to the steady state occurs, a relaxation-oscillation is evident in both the Aegean and Adriatic Seas before arriving to a stable steady state. In this process, under active MOC state, deep layers are getting saltier, leading to a stabilization of the water column and to shutdown or weakening of the MOC. Then, diffusion of heat from the surface to the bottom layers destabilizes it once again and the MOC is activated or getting stronger. A similar mechanism was proposed by Winton and Sarachik (1993) using a box model as well as an OGCM.

Another noticeable finding of this work is the minor to none outflow from both the Aegean and Adriatic Seas in the warm steady state of Exp2. When none of the seas produce dense enough water, the Levantine basin deep layers are not ventilated and a sapropel-like period is enabled (Rohling,
1994). It seems that this lack of ventilation dominates over a wide range of positive temperature anomalies over the Adriatic and further understanding of this phenomenon is needed. The Adriatic passive steady state of Exp1 is also characterized by period of minor outflow from the Aegean, which might explain the presence of those periods.

Although the Aegean MOC is not reversed in the two stable steady states found, each state has different features. Together with a distinct ceasing of the Adriatic MOC in the FGW.m0p2 steady state, our results might explain the occurrence of EMT-like transitions (from almost no dense water outflow to 1 Sv outflow in the Aegean while the Adriatic dense water outflow weakened). Furthermore, the large decadal variability found in this work may be relevant to past climate change occurred in the Mediterranean Sea. This will be explored in a future study.

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References


Manca, B. B., Budillon, G., Scarazzato, P., Ursella, L., 2003. Evolution of dynamics in the eastern mediterranean affecting water mass structures and


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Highlights

• We model the Mediterranean Sea using oceanic general circulation model.
• Eastern Mediterranean overturning circulation multiple equilibria states are found.
• Water exchange between the Aegean and the Adriatic Seas highly affect these states.
• Our results might explain the occurrence of the Eastern Mediterranean Transient.