A process study of the Adriatic-Ionian System baroclinic dynamics

Settore scientifico-disciplinare: GEO/12

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Abstract

The AIS (Adriatic-Ionian System) plays an important role in the Eastern component of MTHC (Mediterranean Thermohaline Circulation). The Adriatic Sea is the main source of Deep Water for the Eastern Basin while the Ionian Sea represents a sort of crossroad point for different kind of waters in the Mediterranean. In the Ionian the bottom layers are filled with dense and oxygenated waters produced in the Adriatic (ADW, Adriatic Deep Water), the intermediate layers are filled with warmer and saltier waters moving westwards and produced in the Levantine basin (LIW, Levantine Intermediate Water). The surface layers are influenced by fresher and colder waters moving eastwards (AW, Atlantic Water). AW occasionally bifurcates northwards, entering the Southern Adriatic.

Recently, the variability of the upper layer circulation in the Ionian Sea, which periodically reverses from anticyclonic to cyclonic and vice versa, has been object of an intense and challenging debate in order to identify the mechanism(s) driving these reversals. It has been suggested that this variability is driven by internal oceanic processes (BiOS, Adriatic-Ionian Bimodal Oscillating System) linking the characteristics of deep water formed in the southern Adriatic with the alternate advection of AW/LIW in this area in correspondence of an anticyclonic/cyclonic pattern in the Ionian. In the BiOS paradigm it has been supposed that the Adriatic (not an external forcing) is the source of potential vorticity/energy driving the circulation of the Ionian. On the other hand it has been suggested that the variability of the circulation in the Ionian is a consequence of changes in the wind stress curl over the basin.

In order to assess the relative importance of remote forcings (wind stress and thermohaline fluxes) on the vorticity and energy budget and to study the baroclinic dynamics of AIS a coarse resolution primitive equation numerical model, based on the MIT general circulation model (MITgcm), is used. The approach followed is based on an increasing complexity in the model forcings and domain. The BCs
at lateral boundary/surface come from both experimental/model datasets and have been chosen carefully in order to reproduce the deep water formation in the southern Adriatic, the main characteristics of the wind field and LIW/AW inflow in the domain.

The approach followed has shown that the influence of the Adriatic Sea outflow on the Ionian Sea dynamics is not affected by the bathymetry of the basin, resulting in both cases in a bi-layer structure for vorticity in the Ionian Sea induced by vortex stretching mechanism. Vorticity and energy balance of AIS appear to be more influenced by the Levantine Basin: this influence is strictly correlated with substantial inflow of Available Potential Energy ($APE_{QG}$) in AIS from the Cretan Passage/Kythira Strait. The characteristics of this inflow strongly depend on the thermohaline properties of waters entering in the region from east. Significant changes in the potential density profiles of these waters mainly at level of surface/intermediate layers (up to 400m) have been observed during the reversal of circulation which took place in the Ionian in 1997, 2006 and 2011. The wind itself appears to have a marginal role in the vorticity/energy budget of the system: it is able to reinforce/weaken the circulation but not to induce changes in sign in the circulation itself.

The final outcomes of this work do not support the idea of a bipole AIS as reported in the BiOS mechanism. On the contrary they support the idea of multipole structure of AIS including the Cretan Passage and the Cretan Sea.

New experiments are planned to include as eastern boundary conditions Temperature/Salinity profiles recently available for Cretan Sea/Passage. Incoming analysis will try to clarify the role of the interannual variability of wind stress forcing in shaping the interaction processes of this multipole structure and the decadal variability of the Ionian upper layer circulation. Another important issue to resolve is the understanding of feedback mechanisms and time delays between the variability of thermohaline properties of Levantine and Ionian waters. Finally, based on most advanced GCM hindcasts for the Mediterranean Sea, an historical analysis will be done in order to identify previous reversals in the Ionian and the characteristics of wind stress/thermohaline fluxes over the area during those periods.
# List of Figures

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1</td>
<td>The coastlines and political geography of the Mediterranean Sea together with the major Sea names [46].</td>
</tr>
<tr>
<td>1.2</td>
<td>Simplified scheme of MTHC with the main water masses involved and main sites of DWF process; redrawn from [27].</td>
</tr>
<tr>
<td>1.3</td>
<td>Scheme of the surface circulation in the Mediterranean Sea from recent observational data and model simulations [46].</td>
</tr>
<tr>
<td>1.4</td>
<td>Simplified scheme of MTHC before/after (a) and during (b) EMT; [61] and references inside.</td>
</tr>
<tr>
<td>1.5</td>
<td>Schematic of EMT-induced flow [50] in the EMED. Triangles indicate head of a streamline; dashed lines indicate less steady flow or spreading. Thinner streamlines indicate relatively lower flow rates. (1) Main route of waters carrying Aegean outflow, fed through the Kasos and Antikithira Straits; later in the EMT, there has been near-bottom counter-flow in the Hellenic Trench. (2) Overflow of EMR during EMT (after 1991). (3) Eastward spreading north and south of the EMR. (4) Flow westward through the Herodotus Trough. (5) Adriatic outflow.</td>
</tr>
<tr>
<td>1.6</td>
<td>Adriatic [32] and Ionian [41] bathymetry.</td>
</tr>
<tr>
<td>1.7</td>
<td>Average ADT in the period July 1993-June 1997 (anticyclonic state in the Ionian) (a) and in the period July 1997-June 2001 (cyclonic state in the Ionian) (b) [5].</td>
</tr>
<tr>
<td>1.8</td>
<td>Schematic representation of the Adriatic-Ionian bimodal oscillating system: (a) anticyclonic mode and (b) cyclonic mode [14].</td>
</tr>
<tr>
<td>1.9</td>
<td>Time series of the average salinity (red dots) and potential density (blue dots) calculated for the 200-800 m layer in the Southern Adriatic. Green dots represent the meridional geostrophic surface current component in the northwestern Ionian. Average horizontal SLA maps for the fall period (October, through December) in the Ionian are shown for 1994 (b), 1995 (c), 1997 (d), 1998 (e) and 2006 (f) [14].</td>
</tr>
<tr>
<td>1.10</td>
<td>Temporal evolution of ADT in the AIS between 1996 and 2013 (Courtesy of G. Civitarese and M. Menna).</td>
</tr>
<tr>
<td>Figure</td>
<td>Description</td>
</tr>
<tr>
<td>--------</td>
<td>-------------</td>
</tr>
<tr>
<td>2.1</td>
<td>The range of phenomena that can be simulated by MITgcm: convection on the left, global circulation patterns on the right [1].</td>
</tr>
<tr>
<td>2.2</td>
<td>Schematic of the integrated continuity equation over the topography [1][10].</td>
</tr>
<tr>
<td>2.3</td>
<td>Schematic of the staggered algorithm where thermodynamic variables are staggered in time with the flow variables [1].</td>
</tr>
<tr>
<td>2.4</td>
<td>Upper surface treatment in the MITgcm: (a) the rigid-lid approximation which imposes a pressure on the fluid, (b) the linear free-surface which permits shallow water waves to propagate [1].</td>
</tr>
<tr>
<td>2.5</td>
<td>Comparison of advection schemes (<a href="http://mitgcm.org/sealion/online_documents/node80.html">http://mitgcm.org/sealion/online_documents/node80.html</a>) in two dimensions; diagonal advection of a resolved Gaussian feature. Courant number is 0.47 with 30×30 points and solutions are shown for $T=\frac{1}{4}$. White lines indicate zero crossings and initial maximum values (ie. the presence of false extrema).</td>
</tr>
<tr>
<td>2.6</td>
<td>Domain of integration in CLNOWIND ideal bathymetry case experiments.</td>
</tr>
<tr>
<td>2.7</td>
<td>Domain of integration in the CLNOWIND real bathymetry case experiments.</td>
</tr>
<tr>
<td>2.8</td>
<td>Domain of integration in OPWIND experiments.</td>
</tr>
<tr>
<td>2.9</td>
<td>Levels spacing in the numerical experiments CLNOWIND/OPWIND-real bathymetry case.</td>
</tr>
<tr>
<td>2.10</td>
<td>MedCordex domain (<a href="http://www.medcordex.eu/cordex_domains_250610.pdf">http://www.medcordex.eu/cordex_domains_250610.pdf</a>).</td>
</tr>
<tr>
<td>2.11</td>
<td>Wind stress curl divided by $\rho = 1028 , \text{kg/m}^3$ and $H=500 , \text{m}$ (in $s^{-2}$) over the AIS: (a) 1987-1995, (b)1996-2012.</td>
</tr>
<tr>
<td>2.12</td>
<td>Wind stress curl divided by $\rho = 1028 , \text{kg/m}^3$ and $H=500 , \text{m}$ (in $s^{-2}$) over the Ionian: (a) 16E-20;32N-40N, (b)16E-20E ; 34N-38N.</td>
</tr>
<tr>
<td>2.13</td>
<td>Monthly Sea Surface Height simulated by MyOcean in October 1995 (a) and in October 2001 (b).</td>
</tr>
<tr>
<td>2.14</td>
<td>Average annual vertical profiles of the zonal component of current in $m/s$ applied to Western Boundary for the period 1987-1995 (a) and 1996-2012 (b).</td>
</tr>
<tr>
<td>2.15</td>
<td>Hovmoeller diagram of zonal component of current in $m/s$ in the Western Boundary corresponding to Sicily channel in the period 1987-1995(a) and 1996-2012 (b).</td>
</tr>
<tr>
<td>2.16</td>
<td>Average annual vertical profiles of the zonal component of current in $m/s$ applied to Eastern Boundary for the period 1987-1995 (a) and 1996-2012 (b).</td>
</tr>
<tr>
<td>2.17</td>
<td>Hovmoeller diagram of zonal component of current in $m/s$ in the Eastern Boundary corresponding to Cretan Passage in the period 1987-1995 (a) and 1996-2012 (b).</td>
</tr>
<tr>
<td>2.18</td>
<td>Hovmoeller diagram of zonal component of current in $m/s$ in the Eastern Boundary corresponding to the entrance of Kythira Strait in the period 1987-1995 (a) and 1996-2012 (b).</td>
</tr>
<tr>
<td>2.19</td>
<td>Spatial coverage of Seasonal Vertical Profiles of Temperature and Salinity of Medar-Medatlas II dataset (<a href="http://doga.ogs.trieste.it/medar/climatologies/medz.html">http://doga.ogs.trieste.it/medar/climatologies/medz.html</a>).</td>
</tr>
</tbody>
</table>
LIST OF FIGURES

2.20 Seasonal vertical profiles of Temperature (in C) (a) and Salinity (in psu) (b) in
the Sicily Channel according to Medar-Medatlas II dataset in winter (red), spring
(blue), summer(green), fall (fuchsia) .................................................. 40

2.21 Seasonal vertical profiles (in red) and median profile (blue) of Temperature (in C)
in the Kythira Strait (a) and Cretan Passage (b) according to Medar-Medatlas II
dataset .............................................. ........................................ 42

2.22 Seasonal vertical profiles (in red) and median profile (blue) of Salinity (in psu)
in the Kythira Strait (a) and Cretan Passage (b) according to Medar-Medatlas II
dataset .............................................. ........................................ 43

2.23 Spatial distribution of the float in the Mediterranean region updated at 22 January
2015 (from http://nettuno.ogs.trieste.it/sire/medargo/active/index.php
(a) and in the area of Eastern Boundary in the period 2005-2013 (b) ............. 45

2.24 Vertical profiles of Temperature (in C) in the Kythira Strait (a) and Cretan Pas-
sage (b) according to Float-Argo ...................................................... 46

2.25 Vertical profiles of Salinity (in psu) in the Kythira Strait (a) and Cretan Passage
(b) according to Float-Argo ...................................................... 47

2.26 Vertical profiles of Salinity (in psu) and Temperature (in C) in the Cretan Passage
(a,b) and Kythira Strait (c,d) according to the data provided by W.Roether ...... 49

2.27 Meridional transport through Otranto Strait according to CLNOWIND experi-
ments (red and blue line) and OPWIND (green line) ............................. 50

2.28 Initial Conditions : vertical profile of Temperature (in C) (a) and Salinity (in psu)
(b) ................................................................. 52
### List of Tables

1.1 Acronyms used in this thesis ........................................ 5

2.1 MITgcm configuration in the experiments .......................... 21

2.2 Name, year and period of the German cruises .................. 48

2.3 Values of diffusivity coefficient for momentum, temperature and salinity ....................................... 51
Objectives of the work

The Adriatic-Ionian System (AIS) is an important component of the Mediterranean Thermohaline Circulation (MTHC): the Adriatic Sea is considered the main source of Eastern Mediterranean Deep Water (EMDW) through deep water formation processes which take place both in its Northern and Southern Basin. The Ionian can be considered a sort of crossroad point for different kind of waters in the Mediterranean basin. Its deep layers are filled with dense and oxygenated waters produced in the Adriatic, its intermediate layers are filled with warmer and saltier LIW (Levantine Intermediate Waters) produced in the Levantine Basin and moving westwards. Its surface layers are influenced by fresher and colder Atlantic Water (AW) coming from Gibraltar strait moving eastwards and occasionally bifurcates northwards entering the Adriatic Sea.

In the last 20 years the variability observed in the upper layer circulation of the Ionian Sea has been the subject of an intense and challenging debate about the identification of the mechanism(s) driving this variability. A negative internal feedback mechanisms (BiOS) between Ionian circulation and thermohaline properties of deep water formed in the Adriatic Sea [8] [14] or a variation in the wind stress curl over the basin [24] [47] have been indicated as possible explanations for the variability observed (for more details see par. 1.2).

The BiOS mechanism suggests that AIS is a bi-modal oscillating system which is able to sustain itself/its circulation through internal processes. Namely, the variation of thermohaline properties of deep waters formed in the southern Adriatic due to alternate advection of LIW/AW in the basin as a consequence of a change in sign of the circulation in the Ionian [8] [14]. On the other hand the second mechanism suggests the idea of a forced mechanism where the source of energy and variability is an external forcing, the wind stress [24] [47].

The aim of this study is to provide a deeper insight into the dynamics of the AIS through a modeling approach, in order to assess the relative impact of external forcing and internal processes acting on the system. We do this by evaluating the vorticity and energy budget of our domain.
A series of numerical experiments have been performed with the following protocol: first we set up an idealized model for the AIS and then we progressively refine the physics and forcing of the model, increasing the complexity of our simulations in order to rank the relevant mechanisms acting on the variability as modeled in our experiments.

This work is organized as follows: chapter 1 provides a brief introduction to Mediterranean Thermohaline Circulation (MTHC), its variability (e.g. Eastern Mediterranean Transient), the dynamics of the Adriatic-Ionian System (AIS) and the mechanisms suggested to explain the variability observed in the Ionian circulation in last 20 years. Chapter 2 provides the description of numerical model and the forcing used in the experiments. Particular attention is given in the description of the numerics of the model, its main parametrizations and the datasets employed as boundary conditions. Chapter 3, in form of draft paper, provides an analysis of the vorticity and an introductive analysis of energy budget of the AIS for a series of experiments with increasing complexity in terms of domain and boundary conditions. Chapter 4 provides the conclusions with some remarks and possible future developments of the work.
Chapter 1

Introduction

This chapter provides a brief introduction to the state of art on Mediterranean Thermohaline Circulation (MTHC) and the dynamics of the Adriatic-Ionian system.

1.1 The Mediterranean Thermohaline Circulation (MTHC)

The Mediterranean Sea (Figure 1.1) is a semi-enclosed basin located at mid-latitudes between Africa and Europe communicating with the World Ocean through the narrow and shallow strait of Gibraltar [56] [61]. It is characterized by a complex land-sea distribution and is divided in two sub-basins, namely Eastern Mediterranean (EMED) and Western Mediterranean (WMED, hereafter for the acronyms in the text refer to Table 1.1) communicating through the Sicily Channel [56] [61]. It is a region where oceanic processes such as Deep Waters Formation (DWF) occur at a smaller scale resulting in a basin-wide MTHC (Figure 1.2) [56] [61].

MTHC is driven by the buoyancy fluxes at the surface of Mediterranean Sea and it is an open cell which starts from Gibraltar strait with an inflow of relatively fresh water (AW) and ends in the same site with a undercurrent of LIW, saltier and warmer [56] [61] (Figure 1.2). MTHC exhibits a strong seasonal and interannual variability and has a complex structure, involving different water masses with different Temperature-Salinity characteristics [56] [61].

After entering the Mediterranean through Gibraltar Strait AW moves eastwards (Figure 1.2, Figure 1.3) along the Northern African coastlines forming the Algerian current [36] [38] [39] [56]. This current is unstable and gives rise to meanders and coastal eddies of 50-100 km of diameter [56] [38] [39] [56]. Then it divides in two branches, one entering the Tyrrenian sea and the other continuing through Sicily Channel (Figure 1.2, Figure 1.3) [56] [56]. The branch entering the Tyrrenian sea moves along the Sicily coastline, then along the Western Italian coastlines, Ligurian, Provencal and Catalan coastlines forming a complex system of surface currents (Figure 1.3) [56] and references inside [37], [61] and reference inside).

The main pathway of the AW crossing the Sicily channel and moving eastwards is still a
Figure 1.1: The coastlines and political geography of the Mediterranean Sea together with the major Sea names [46].

Figure 1.2: Simplified scheme of MTHC with the main water masses involved and main sites of DWF process; redrawn from [27].
<table>
<thead>
<tr>
<th>Acronym</th>
<th>Meaning</th>
</tr>
</thead>
<tbody>
<tr>
<td>MHTC</td>
<td>Mediterranean Thermohaline Circulation</td>
</tr>
<tr>
<td>WMED</td>
<td>Western Mediterranean</td>
</tr>
<tr>
<td>EMED</td>
<td>Eastern Mediterranean</td>
</tr>
<tr>
<td>DWF</td>
<td>Deep Water Formation</td>
</tr>
<tr>
<td>AW</td>
<td>Atlantic Water</td>
</tr>
<tr>
<td>MMJ</td>
<td>Mid Mediterranean Jet</td>
</tr>
<tr>
<td>LSW</td>
<td>Levantine Surface Water</td>
</tr>
<tr>
<td>LIW</td>
<td>Levantine Intermediate Water</td>
</tr>
<tr>
<td>WMDW</td>
<td>Western Mediterranean Deep Water</td>
</tr>
<tr>
<td>EMDW</td>
<td>Eastern Mediterranean Deep Water</td>
</tr>
<tr>
<td>EMT</td>
<td>Eastern Mediterranean Transient</td>
</tr>
<tr>
<td>AIS</td>
<td>Adriatic-Ionian System</td>
</tr>
<tr>
<td>ADW</td>
<td>Adriatic Deep Water</td>
</tr>
<tr>
<td>RG</td>
<td>Rhodes gyre</td>
</tr>
<tr>
<td>SAG</td>
<td>South Adriatic Gyre</td>
</tr>
<tr>
<td>SHG</td>
<td>Shikmona Gyre</td>
</tr>
<tr>
<td>NIG</td>
<td>Northern Ionian Gyre</td>
</tr>
<tr>
<td>CDW-CSOW</td>
<td>Cretan Deep Water</td>
</tr>
<tr>
<td>BIOS</td>
<td>Adriatic-Ionian Bimodal Oscillating System</td>
</tr>
</tbody>
</table>

Table 1.1: Acronyms used in this thesis

Figure 1.3: Scheme of the surface circulation in the Mediterranean Sea from recent observational data and model simulations [46]
subject of debate ([56] and references inside). According to in-situ and mean dynamic topography data ([31], [56] and references therein) AW has two branches: one moving clockwise in the Northern Ionian and then crossing the Cretan Passage in form of cross-basin jet (MMJ), another spreading directly along the Libyan and Egyptian coastlines ([56] and references inside). On the other hand satellite data ([49], [56] and references inside) suggest an eastward along-slop flow moving continuously along the African coastlines.

The progressive increase of salinity due to evaporation and mixing with higher salinity waters constitutes a preconditioning for the production of high-density waters through winter cooling, entrainment processes and further evaporation (Figure 1.2, [61]). The LSW (AW modified through evaporation in the Eastern Basin) reaches salinities around 38.9 psu. LSW is interested by further increased of density through evaporation and cooling during winters and becomes the main source, in the region of the Rhodes gyre (Figure 1.3, [49], [23], [61]) of LIW ([61] (Figure 1.2). LIW occupies the intermediate layers of Mediterranean Sea (between 200 m and 500 m [26], [61]) with a salinity maximum of 38.95 - 39.05 psu (Figure 1.4a) [61] near the production areas which reduces progressively to 38.4 psu at the Mediterranean outflow at the Gibraltar strait [3], [5], [6], [60], [61]. The LIW flows westwards with a branch entering the Ionian Sea and then the Sicily strait, finally exiting at Gibraltar strait [61] (Figure 1.2), and another branch entering the Adriatic Sea [56].

DWF occurs both in WMED and EMED (Figure 1.2, Figure 1.4a) [56]. In the WMED (Figure 1.2, Figure 1.4a) DWF take place in the Gulf of Lions where penetrative open-ocean convection exerts an energetic mixing between AW and LIW all along the water column, sometimes reaching the bottom, leading to formation of WMDW ([57], [35], [61], [56] and reference inside) which spreads out from the convective region filling the bottom of WMED. In the EMED (Figure 1.4a) both Aegean sub-basin [25], [40] and Adriatic Sea [29], [55] are recognized as a potential contributor to the waters filling the deep Ionian and Levantine basins (EMDW [61], [56] and reference inside) despite the deep waters formed in the Aegean sub-basin are not enough dense to reach the bottom layers of the Ionian. The Adriatic Sea is considered as the major source of the EMDW (hereafter called ADW) [29], [55], [61] and reference inside). Here DWF (see for more details par 1.2) takes place both in the shallow Northern Adriatic and the deep Southern Adriatic [3], [55].

A large change of DWF in the EMED (Figure 1.4b) occurred during the 1990s, as revealed by a comparison of Temperature and Salinity data collected during the years 1987 and 1995 by the German Cruises M5/6 and M31/1 [61], [51], [22], [27], [29], [50], [48], [59]. These observations showed that the main DWF in EMED shifted from the Adriatic to the Aegean Sea. This event has been called Eastern Mediterranean Transient (EMT) [61], [51], [22], [27], [29], [50], [48], [59]. During EMT warmer (13.7°C), highly saline (38.8 psu) and rich in oxygen waters with respect those formed in the Adriatic (13.3°C and 38.68 psu) and originated from the Straits of the Cretan Arc (Fig-
Figure 1.4: Simplified scheme of MTHC before/after (a) and during (b) EMT ([61] and references inside)
ure [1.4.b] filled the bottom layers of the Ionian and Levantine basins. EMT started since 1987 with the increase of densities of deep waters formed in the Aegean sub-basins as a consequence of an excessive winter cooling during the winter 1987-1993 [55] that acted on top of a previous salinity increase taking place between 1987 and 1991 [27] [50]. These densities values reached a maximum value during the 1993 (\(\sigma_2=37.96 \ \text{kg} \ \text{m}^{-3}\) at 900 m; \(\sigma_o=22.35 \ \text{kg} \ \text{m}^{-3}\)) and then decreased gradually [50]. The principal outflow route (Figure 1.5) for these dense waters was represented by the Kasos Strait and they started to affect the deep layers of EMED about 1990 with a peak in the delivering outflow around 1994 [50]. The mean value of the outflow have been estimated about 3 Sv (mid 1992-1994) and the total amount was about twice the total volume of the Aegean Sea [50] [59]. The Aegean contributions exiting via Kasos Strait propagated westwards adjoining the Cretan Arc, mostly continuing up to the northern end of the Hellenic Trench near 37 N [50] (Figure 1.5). Then the waters encountered fast spreading across the Ionian by circulation (partly cyclonic along the periphery of the Ionian Sea) and mixing (Figure 1.5). After 1991, Kasos outflow gradually expanded also into the southeastern Levantine Sea (Figure 1.5) [50]. The influence of Aegean waters on temperature and salinity, most on the Levantine Sea and western Hellenic Trench [50], shrunk with time. EMT event is relaxed by about 1995 [59] and the outflow almost stopped in 1998 [23], although the densities in the Cretan Sea in the 2001 were still high enough to sustain further outflow into the EMED [50]. It appears now that the main contribution of EMDW is the Adriatic again [23].

Several scenarios have been offered in order to provide insight to the EMT causes such as:

- Local (over the southern Aegean) anomalous meteorological conditions [27]
- Changes in the mesoscale circulation in the Ionian Sea [29]
- Reduced Black Sea outflow [62]
- Anomalously high buoyancy losses from the Aegean [21]
- Damming of main rivers in Egypt and the ex-USSR [9]
- Wind Anomalous Field [54]

\(\sigma_2\) is the potential density referenced to 2000 dbar pressure while \(\sigma_o\) is the potential density referenced to the surface.
1.2 The Adriatic-Ionian System (AIS)

In the previous section, the importance of AIS in the MTHC was introduced. The Adriatic Sea (Figure 1.6a) is a semi-enclosed elongated basin (800 km long and 200 km wide), which communicates with the Ionian Sea through the Otranto Strait (75 km wide and up to 800 m deep) [20] [32]. The basin can be separated from north to south into three sub-basins [32]:

- Northern Adriatic (NA) which includes the very shallow parts of the basin in the north and extends southwards to the 100m
- Middle Adriatic (MA), in the central part of the basin, which includes the Middle Adriatic Pit and extends southwards to the Pelagosa Sill.
- The Southern Adriatic (SA) extends from this sill to the Otranto strait, with bathymetry that reaches 1200m at the bottom of a wide depression at its center, known as the Southern Adriatic Pit (SAP).

Two types of DWF occur during winter within the Adriatic Sea [61] [32] [20] and are facilitated by:

- The cold and dry Bora wind blowing over the region during winter and inducing large buoyancy losses at the surface [32].
the presence of the LIW at its intermediate layers [32, 20].

The major portion of the ADW (about 82% [32]) is formed through open convection inside the SAP within the cyclonic gyre of the SA [61, 32, 20], while the remaining part has its origin on the continental shelf of the NA and MA [61]. In the SA the circulation is permanently cyclonic and the occasional intrusion of LIW through Otranto Strait enhances the potential of dense water production [32, 20]. The dense waters produced over the NA by shelf formation propagate to the south and mix with the dense water produced in the deep southern Adriatic [32, 20]. The ADW overflows as an undercurrent through the Otranto Strait oxygenating the deep layers of the Ionian Sea and EMED. The annual export of ADW towards the Ionian Sea, measured throughout the period 1997 to 1999, ranged between 0.1 Sv in 1997 to 0.4 Sv in 1999 ( [17, 31, 61] and reference inside).

The Ionian Sea (Figure 1.6b) is the deepest basin of the Mediterranean Sea. It is connected with the WMED through the Sicily channel (130km wide), with the Adriatic Sea through the Otranto strait, with the Aegean Sea through Kythira/Antikythira strait (70km wide) and with the Levantine Basin through the Cretan Passage. As previously observed, it plays an important role in the MTHC (par 1.1) [14]. ADW spreads into the Ionian bottom layer, whilst the intermediate layer is influenced by salty LIW. Furthermore, through the Sicily channel, the AW enters the Ionian, propagates toward the Levantine basin and, occasionally, bifurcates northwards [14].

The EMED and the Ionian are characterized by the presence of a number of topographic (or bottom-trapped) often wind-induced gyres such as Rhodes gyre (RG), Shikmona gyre (SHG), Southern Adriatic Gyre (SAG), Northern Ionian Gyre (NIG). . . ) [56]. These gyres are characterized by a constant and well defined vorticity ([56] and references therein). The NIG appears to be an exception to this rule: a change in rotation of NIG takes place on decadal scales and it is able to affect the spreading of AW and LIW in the Ionian ([56] and reference inside):

- In a cyclonic state, AW flows eastwards along the African coast. In the anticyclonic state, AW propagates northwards, eventually entering the Adriatic [56]. The eastwards flow of AW through the Cretan Passage is weakened leading to an increase of salinity in Levantine Basin and in the corresponding LIW [15].

- When NIG is cyclonic, the LIW flow, through Cretan Passage is intensified, when NIG is anticyclonic both AW and LIW are weakened [56].

This variation in NIG sign rotation is able to condition the thermohaline properties of Adriatic, Ionian [13] and Levantine Basins [15], the salt content in the Sicily channel [22], the DWF in the Adriatic and Aegean Sea [56] but also the amount of nutrients imported into the Adriatic from the Ionian [13] and the biodiversity patterns [13]. The variation on the decadal scale appears to be confirmed by the occurrence of several high salinity events in the Adriatic [56] [14].
Figure 1.6: Adriatic \cite{32} and Ionian \cite{41} bathymetry
The reversal in the sign of the NIG has been often associated with changes in wind stress curl ([24] [30] [59] and reference inside). On the other hand, the first reversal (from anticyclonic to cyclonic pattern) observed in the middle of 1997 (Figure 1.7) [8] [14] has been attributed to the relaxation of EMT [8]. In fact before 1997, CDW (former CSOW) occupied mainly the eastern and northeastern flanks of the Ionian and its concentration decreased moving westwards. This resulted in a bottom pressure gradient directed towards the center of the basin, which sustained a stationary cyclonic shear in the bottom layer and the anticyclonic one in the upper thermocline layer (Figure 1.7a) [8]. After 1997 (Figure 1.7b), CDW reaching the northwestern Ionian was mixed with the ADW (less dense than CDW) in the central abyssal portion of the basin setting up a bottom pressure gradient directed from the center of the basin toward the coast, sustaining a stationary anticyclonic shear in the bottom layer and the cyclonic one in the upper thermocline layer [8]. Comparing the rate of change of the vorticity [14] due to internal processes (mainly stretching) and the source of vorticity due to the wind stress, it has been suggested that these variations in the Ionian circulation can be effectively driven by internal oceanic processes, which can outweigh wind stress [8] [14]. These processes have been suggested to be a result of feedback mechanism between the redistribution of water masses in the Ionian and the thermohaline properties of deep waters formed in the Southern Adriatic [14].

This feedback mechanics has been called BiOS (Figure 1.8) [14]. In the BiOS formulation:

Figure 1.7: Average ADT in the period July 1993-June 1997 (anticyclonic state in the Ionian) (a) and in the period July 1997-June 2001 (cyclonic state in the Ionian) (b) [8]
Figure 1.8: Schematic representation of the Adriatic-Ionian bimodal oscillating system: (a) anticyclonic mode and (b) cyclonic mode [14]

Figure 1.9: Time series of the average salinity (red dots) and potential density (blue dots) calculated for the 200-800 m layer in the Southern Adriatic. Green dots represent the meridional geostrophic surface current component in the northwestern Ionian. Average horizontal SLA maps for the fall period (October, through December) in the Ionian are shown for 1994 (b), 1995 (c), 1997 (d), 1998 (e) and 2006 (f) [14]
• the Adriatic-Ionian system behaves as a bimodal oscillating system

• when anticyclonic circulation is present in the Ionian, AW are deflected in the Adriatic leading to production of ADW of lowering density which spread in the Ionian producing a deepening of isopycnal surface and stretching of water column resulting in a reversion from anticyclonic to cyclonic (Figure 1.8, Figure 1.9).

• when cyclonic circulation is present in Ionian saltier LIW enters the Adriatic leading the production of ADW of increasing density which produce a shallowing of isopycnal surface, leading to a reversion from cyclonic to anticyclonic (Figure 1.8, Figure 1.9).

Altimetric maps (Figure 1.10) showed an anticyclonic mode starting in 2006, a cyclonic mode starting in 2011 and unexpectedly ending in 2012 when a new reversal to anticyclonic mode has been observed. This premature inversion has been attributed to extremely strong winter in 2012, which caused the formation of very dense ADW, flooding Ionian flanks in May and inverting the bottom pressure gradient [16].

Figure 1.10: Temporal evolution of ADT in the AIS between 1996 and 2013 (Courtesy of G. Civitarese and M. Menna)
Chapter 2

Numerical Model and Forcings data

This chapter provides a description of the numerical model and the data employed in our numerical experiments.

2.1 Numerical Model: MIT General Circulation Model (MITgcm)

The dynamics of AIS in response to some forcings have been simulated with MITgcm (MIT general circulation model, [1] and references inside, [2]), a primitive equations model. MITgcm\(^1\) is widely used in the scientific community because of its flexibility and its ability to simulate different phenomena at different complexity ([1] and references inside, [2]). In particular:

- it is written in Fortran77
- it can be used to study both atmospheric and oceanic phenomena: in past, present and future conditions
- it can be used to study both small-scale and large scale processes (from convection to ocean circulation, Figure 2.1)
- it employs a finite volume and partial cells techniques which allows the treatment of irregular/complex geometries
- it is developed to perform efficiently on a wide variety of computational platforms

\(^1\)http://mitgcm.org
• it can run in serial/parallel configuration.

Figure 2.1: The range of phenomena that can be simulated by MITgcm - convection on the left, global circulation patterns on the right [1]
2.1.1 Model equations and their discretization

Assuming that the flow is incompressible and the Boussinesq approximation holds the equations in which MITgcm is rooted are the following [1] [33] [34]:

- Momentum:
  \[ \rho_0 D_t \vec{v} + 2\omega \wedge \rho_0 \vec{v} + g\rho_0 k + \nabla p = F \]  \hspace{1cm} (2.1)

- Continuity:
  \[ \rho_0 \nabla \cdot \vec{v} = 0 \]  \hspace{1cm} (2.2)

- Mass Conservation:
  \[ \partial_t \eta + \nabla \cdot (H + \eta) \vec{v}_h = P - E + Runoff \]  \hspace{1cm} (2.3)

- Energy Conservation:
  \[ D_t \theta = Q_\theta \]  \hspace{1cm} (2.4)

- Salt Conservation:
  \[ D_t S = Q_S \]  \hspace{1cm} (2.5)

- State equation:
  \[ \rho = \rho(S, p, \theta) \]  \hspace{1cm} (2.6)

where \( \vec{v} \) is the three dimensional velocity vector, \( p \) is pressure, \( \rho \) is in-situ density, \( \eta \) is the displacement of the free-surface from the resting sea-level, \( \theta \) is the potential temperature, \( S \) is salinity, \( \rho_0 \) is a constant reference density, \( g \) is the constant gravitational acceleration and \( H \) is a fixed-in-time bottom depth [1] [33] [34]. \( F, P-E+Runoff, Q_\theta \) and \( Q_S \) are forcing fields [1].

The model equations are discretized using the finite volume methodology [1] [2] in which the governing equations are integrated over space-filling finite volumes (cells) that make up a discrete grid. The shape of these cells is determined by “intersecting boundary method”. This method allows the boundary which represents the topography to intersect a grid of cells modifying the shape of those cells intersected. The possible final representations of the topography are:

- piecewise linear representation (shaved cells)
- piecewise constant representation (partial cell)

For example integrating (Figure 2.2) the equation (2.2) over a finite volume and applying the Gauss-divergence theorem we get:

\[ A_{east}^u u_{east} - A_{west}^u u_{west} + A_{north}^v v_{north} - A_{south}^v v_{south} + A_{up}^w w_{up} - A_{down}^w w_{down} = 0 \]  \hspace{1cm} (2.7)
where $A$ is the area of a cell face and the budget is written in terms of the normal flow across the cell face. The horizontal components of velocity are staggered on an Arakawa-C grid and the vertical velocity component in a Lorentz grid [1] [2].

![Schematic of the integrated continuity equation over the topography](image)

Figure 2.2: Schematic of the integrated continuity equation over the topography [1] [10]

A project method is used to solve the discretized equations [1] [34]. In the case of eq (2.1) discretizing between time levels $n$ and $n+1$ we get:

$$
\rho_0 \vec{v}^{n+1} + \Delta t \nabla p = \rho_0 \vec{v}^n + \Delta t G = \rho_0 \vec{v}^n
$$

(2.8)

where $G$ includes all the discretized terms of momentum equations excluded the pressure term. Substituting (2.8) in the (2.2):

$$
\delta_i (A^u u^{n+1}) + \delta_j (A^v v^{n+1}) + \delta_k (A^w w^{n+1}) = 0
$$

(2.9)

we find:

$$
\frac{\Delta t}{\rho_0} \left[ \delta_i \left( \frac{A^u}{\Delta x} \delta_x p \right) + \delta_j \left( \frac{A^v}{\Delta y} \delta_y p \right) + \delta_k \left( \frac{A^w}{\Delta z} \delta_z p \right) \right] = \delta_i (A^u u^*) + \delta_j (A^v v^*) + \delta_k (A^w w^*)
$$

(2.10)

where $u^* = u^n + \Delta t G_u$, $v^* = v^n + \Delta t G_v$ and $w^* = w^n + \Delta t G_w$. (2.10) is a three dimensional elliptic equation particularly expensive to solve [1]: in the model this problem is overcome through a decomposition of pressure in three dynamical parts [1] [34]:

$$
p = p_s(x, y) + p_h(x, y, z) + p_{nh}(x, y, z)
$$

(2.11)

where $p_s$ is the pressure imposed by rigid-lid (or free surface which requires a different numerical treatment) and is constant [34], $p_h$ is hydrostatic pressure and $p_{nh}$ non-hydrostatic pressure.

The $p_s$ equation is found neglecting in the eq. (2.11) the term associated with gradient of non-hydrostatic pressure $p_{nh}$:

$$
\nabla_z \cdot H \nabla_z p_s = \frac{\rho_0}{\Delta t} \nabla \cdot \int_{-H}^{0} (\vec{v}_h^* - \Delta t \nabla_z p_h) dz
$$

(2.12)
Once found the $p_s$ the (2.11) is solved for $p_{nh}$:

$$\nabla^2 p_{nh} = \frac{p_0}{\Delta t} \nabla \cdot \left( \vec{v}^n - \Delta t \nabla_z p_s - \Delta t \nabla_z p_h \right) \quad (2.13)$$

The advantage of this decomposition is that the surface pressure equation is only a two-dimensional elliptic equation and thus much cheaper to solve. Thermodynamics variables are integrated forward in time using a staggered algorithm (Figure 2.3) [34] [2]:

- $\theta^{n+\frac{1}{2}} = \theta^{n-\frac{1}{2}} + \Delta t (Q^0_\theta - \nabla \cdot F(\vec{\theta}^n, \tilde{\theta})) \quad (2.14)$

- $S^{n+\frac{1}{2}} = S^{n-\frac{1}{2}} + \Delta t (Q^0_s - \nabla \cdot F(\vec{\theta}^n, \tilde{S})) \quad (2.15)$

- $\rho^{n+\frac{1}{2}} = \rho(S^{n+\frac{1}{2}}, \theta^{n+\frac{1}{2}}, p_0(z)) \quad (2.16)$

- $p_h^{n+\frac{1}{2}} = - \int_0^z g \rho^{n+\frac{1}{2}} dz' \quad (2.17)$

- $\vec{v}^n = \vec{v}^n + \Delta t (G(\vec{v}) + F^n + \frac{1}{2} - \nabla_z p^{n+\frac{1}{2}}) \quad (2.18)$

- $\nabla^2 p_{nh} = \nabla \cdot \vec{v}^n \quad (2.19)$

- $\vec{v}^{n+1} = \vec{v}^n - \Delta t \nabla p_{nh} \quad (2.20)$

Staggering allows to treat the gravity wave terms as centered in time and has the stability of a leap-frog scheme for those modes but at half of the cost [1] [34] [2].
Table 2.1 reports the main configuration of the model in all our experiments. MITgcm is able to solve momentum equations in both Hydrostatic and Non-Hydrostatic configuration. The Hydrostatic configuration has been employed in our experiments. MITgcm allows three different treatments of the upper boundary (ocean surface) for a domain [10]. The Rigid-Lid and Linear Free surface have been considered here (Figure 2.4, Table 2.1). The Rigid-lid approximation (Figure 2.4, a) consists essentially in representing the upper surface as an impermeable boundary exerting a pressure on the fluid such that the depth integrated divergence of the volume flux $\nabla \cdot \int_{-H}^{0} \vec{v} dz$ is zero [10].

In the Linear Free surface approach (Figure 2.4, b) some small terms are ignored in the depth integrated continuity equation $\partial_t \eta + \nabla \cdot \int_{-H}^{0} \vec{v} dz = P - E + Runoff$ allowing surface gravity waves to propagate with finite phase speed. This is a very good approximation in deep waters where $\eta \ll H$ [10].

Concerning the vertical column water instability a simple approach based on the convective adjustment algorithm has been used (Table 2.1). This algorithm "scans" the water column and when it finds an unstable density profile between two adjacent layers $z_1 < z_2$ with densities $\rho_1(z_1) > \rho_2(z_2)$ it instantaneously mixes the two layers. The model considers more advanced schemes for vertical mixing such as KPP (K-Profile Parameterization). This scheme has not been employed in our experiments because of some troubles with the numerical stability of the model. Still due to the coarse resolution of the domain (see par. 2.2.1) a constant horizontal/vertical eddy diffusivity for Momentum, Temperature and Salinity have been employed (the values are reported in the tab. 2.3).
Figure 2.4: Upper surface treatment in the MITgcm: (a) the rigid-lid approximation which imposes a pressure on the fluid, (b) the linear free-surface which permits shallow water waves to propagate. 

<table>
<thead>
<tr>
<th>Model configuration</th>
<th>Hydrostatic</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper Boundary Parametrization</td>
<td>Rigid-Lid/Linear Free Surface</td>
</tr>
<tr>
<td>Convective parametrization</td>
<td>Convective Adjustment</td>
</tr>
<tr>
<td>Type of grid</td>
<td>Spherical Polar Grid</td>
</tr>
<tr>
<td>State equation</td>
<td>JMD95Z</td>
</tr>
<tr>
<td>Advection Scheme</td>
<td>33</td>
</tr>
</tbody>
</table>

Table 2.1: MITgcm configuration in the experiments

Many different grid types such as cartesian coordinates, spherical polar coordinates (regular longitude-latitude) and general curvilinear orthogonal coordinates are available: in our present configuration a spherical polar coordinates has been implemented. MITgcm allows the user to specify the preferred equation of state for the sea water. The JMD95Z state equation has been used.

The tracer equations are discretized using the finite volume method described earlier. The default advection scheme for the tracer equations is the centered second order method which requires a second order or quasi-second order time-stepping scheme (as the quasi-second order Adams-Bashforth method, ABII) to be stable. Besides the default scheme, others schemes are available. There are no general rules in the choice whether a scheme is suitable or not for an experiment. In general this choice depends strongly on temporal and spatial resolution, i.e. Courant number. Figure 2.5 shows two dimensional advection of a generic gaussian feature. The left column shows the second order schemes: centered second order with Adams-Bashforth (top),
Lax-Wendroff (middle) and Superbee flux limited (bottom). Middle column shows the third order schemes: upwind biased third order with Adams-Bashforth (top), third order direct space-time method (middle) and the same with flux limiting (bottom). The right column shows the centered fourth order scheme with Adams-Bashforth (top) and a fourth order variant on the DST method (middle) and the superbee flux limiter (second order) applied independently in each direction (method of lines, bottom). Based on previous experiences (Querin S., personal communication) in order to avoid an anomalous and noisy advection of the tracers all the experiments have been run with 3rd order DST scheme (Advection scheme 33, Table 2.1).

Figure 2.5: Comparison of advection schemes [http://mitgcm.org/sealion/online_documents/node80.html] in two dimensions; diagonal advection of a resolved Gaussian feature. Courant number is 0.47 with $30 \times 30$ points and solutions are shown for $T = \frac{1}{2}$. White lines indicate zero crossings and initial maximum values (ie. the presence of false extrema)
2.2 Boundary Conditions

2.2.1 Sets of the experiments

The protocol established for this work has led to prepare a series of experiments with increasing complexity in terms of domain of integration and Boundary Conditions of the model. Two main sets of experiments are considered in this work:

- **CLNOWIND**: closed boundaries for the domain of the integration and no wind forcing applied
- **OPWIND**: open boundaries for the domain of the integration and wind forcing applied

**CLNOWIND**

- Closed Boundaries Domain
- Idealized bathymetry/Real bathymetry
- $0.25^\circ/0.125^\circ$ horizontal resolution
- 24 levels/32 levels vertical resolution
- No Wind Forcing
- Constant Horizontal/Vertical Diffusivity coefficients for Momentum/Temperature/Salinity
- Heat Fluxes over the SA in January-February-Zero for the rest of the year all over the domain
- Timestep 1200s /1100s
- Frequency of output: 432000s
- Length of Simulation: 27 years

**OPWIND**

- Open Boundaries Domain
- Real bathymetry
- $0.125^\circ$ horizontal resolution
- 32 levels vertical resolution
- Wind Boundary Conditions for the period 1987-2012

---

5the period of forcing covers the EMT event and the reversals observed in the Ionian circulation
• Momentum Boundary Conditions for the period 1987-2012
• Temperature and Salinity Boundary Conditions for the period 1987-2012
• Constant Horizontal/Vertical Diffusivity coefficients for Momentum/Temperature/Salinity
• Heat Fluxes over the SA in January-different from zero for the rest of the year all over the domain
• Timestep 1000s
• Frequency of output : 2635200s
• Length of simulation : 25 years

In the next sections we will discuss briefly the main characteristics of the domain of integration for CLNOWIND and OPWIND experiments and the datasets used as boundary conditions.

2.2.2 Domain of integration

The series of experiments are characterized by an increasing resolution, dimension and complexity in the domain. The NA and the Middle Adriatic have not been included in our domains. In the first CLNOWIND experiments an idealized bathymetry (Figure 2.6) has been considered with an Ionian characterized by a flat bottom and 2000 m deep. The Adriatic, in this configuration, is 800 m deep. The domain covers approximately the geographical domain 16E-20E; 31N-42.25N. The two basins are connected through the Otranto Strait. Totally the domain extends over 17 × 49 points with a horizontal resolution of 0.25° in longitude-latitude.

Figure 2.6: Domain of integration in CLNOWIND ideal bathymetry case experiments

24
In the following CLWIND experiments we have introduced a real bathymetry with an increased spatial resolution of 0.125° in longitude-latitude (Figure 2.7, Figure 2.8). This bathymetry has been derived from a higher resolution bathymetry used in the Mediterranean Forecasting System [42]. The maximum depth has been set to 4000 m. The domain (Figure 2.7) extends horizontally over 96 × 96 points (covering the geographical domain of approximately 13E-24.5E and 31N-42.5N). In the case of OPWIND (Figure 2.8), the horizontal extension is equal 108 × 100 points covering approximately the geographical domain 12.4E-25.5E and 30.1N-42.5N. The domain has two open boundaries: a Western Boundary connecting the Sicily with the Northern African Coastlines and corresponding to the Sicily channel and an Eastern Boundary constitutes of two smaller open boundaries: one connecting Creta island with Northern African (hereafter named Cretan Passage) and another forming a sort of channel connecting the Arc of Creta with the Ionian Sea through the Kythira Strait (hereafter called Kythira Strait).

Figure 2.7: Domain of integration in the CLNOWIND real bathymetry case experiments

The vertical resolution of the domain in the case of the experiments with real bathymetry is reported in Figure 2.9. The density of levels is higher in the upper part of the water column while it decreases moving down. In the case CLNOWIND experiments with idealized bathymetry the levels spacing is the same as Figure 2.9 in the first 2000 m.
Figure 2.8: Domain of integration in OPWIND experiments

Figure 2.9: Levels spacing in the numerical experiments CLNOWIND/OPWIND-real bathymetry case
2.2.3 Boundary conditions Datasets

The atmospheric and oceanic boundary conditions employed for our numerical experiments come from two main kind of datasets:

- Model Boundary Conditions
- Experimental Boundary Conditions

Model Boundary Conditions

Zonal and meridional component of wind speed (at 10m) for the AIS have been derived from the RegCM4 simulations carried out in the MedCordex domain (Figure 2.10). This domain has been set for the Med-Cordex project [6,7]. The Cordex project [7] has the final purpose to evaluate the Regional climate downscaling techniques in downscaling global climate projections and quantifying the performance of Regional climate models such as RegCM4.

RegCM4 [8] (The Regional Climate Model system v4) is the latest version of RegCM1, originally developed in 1989 at the National Center for Atmospheric Research (NCAR), and is currently maintained in the Earth System Physics (ESP) section of the ICTP [32].

The dataset derived from the RegCM4 has a spatial resolution of 50km and covers the period 1987-2012: the data which belonged to AIS area have been extracted and interpolated to a regular longitude-latitude grid. Finally a climatology of 12 months has been computed for the

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period 1987-1995 and for the period 1996-2012. Figure 2.11 a, b show the annual mean contribution in term of vorticity variation of wind stress to AIS. In both cases it is clear that AIS has a dipole structure with the Southern Adriatic characterized by the strong cyclonic input all over the year while the Ionian (mainly its central and meridional part) is interested by a permanent anticyclonic input.

Figure 2.12 a, b shows the average variation of vorticity induced by wind stress all along the year in the period 1987-1995 (first 120 months) and 1996-2012 (121-300 months) over the Ionian basin (16E-20;32N-40N) (a) and its central part (16E-20E ; 34N-38N) (b) . Both graphs show that the variation of vorticity is cyclonic over the basin mainly in winter and fall when is most intense the synoptic activity due to cyclones over the region while it is anticyclonic in spring and summer due to the prevalence of anticyclonic conditions. In general the average variation of vorticity induced by the wind stress over the Ionian Sea ranges between $-0.2 \times 10^{-13}\text{s}^{-2}$ and $-0.3 \times 10^{-13}\text{s}^{-2}$.

\footnotetext{the reason for the choice of these periods will be explained in the next section}
Figure 2.11: Wind stress curl divided by $\rho = 1028 \frac{kg}{m^3}$ and H=500m (in $s^{-2}$) over the AIS: (a) 1987-1995, (b) 1996-2012
Figure 2.12: Wind stress curl divided by $\rho = 1028 \text{ kg m}^{-3}$ and H=500m (in s$^{-2}$) over the Ionian: (a) 16E-20:32N-40N, (b) 16E-20E; 34N-38N
Due to a lack of data for the zonal and meridional component of current mainly in the Levantine Basin, the use of model dataset for these variables was necessary. The zonal and meridional component of current used for both open boundaries came from the MyOcean dataset\[10\].

MyOcean is part of the Mediterranean Forecasting System that is a hydrodynamic model based on primitive equation in spherical coordinate (NEMO) implemented in the Mediterranean at 1/16\(^0\) X 1/16\(^0\) horizontal resolution and 72 unevenly spaced vertical levels \[42\]. The domain of integration is the Mediterranean Basin (-6W-36.25E; 30.187N-45.937N) and also extends into the Atlantic in order to better resolve the exchanges with the Atlantic Ocean at the Strait of Gibraltar while the Dardanelles inflow is parameterized as a river. The model is forced by momentum, water and heat fluxes interactively computed by bulk formulae using the 6h, 0.75\(^0\) horizontal-resolution ERAInterim reanalysis fields from the European Centre for Medium-Range Weather Forecasts (ECMWF).

A variational data assimilation scheme (OceanVAR) is employed and the assimilated data include: sea level anomaly, sea surface temperature, in situ temperature and salinity profiles by argo floats, and in situ temperature and salinity profiles from CTD \[42\]. The model has been initialized at the 1st January 1985. The assimilation of the available satellite and in situ data is done since January 1st 1985 too. Two year of spin-up are considered, thus the available data starts in 1987. The outputs of the model include:

- zonal sea water velocity
- meridional sea water velocity
- sea surface height above sea level
- sea water potential temperature
- sea water salinity

Figure 2.13 shows the monthly Sea Surface Height in the October 1995 (a) and 2001(b).

The monthly data of MyOcean covers the period 1987-2012: the vertical profiles of velocities corresponding to Western and Eastern Boundary of our domain have been extracted and then interpolating to a lower resolution along the horizontal dimension and along our levels for the vertical dimension. Also in this case a climatology of 12 months has been computed for the period 1987-1995 and the period 1996-2012. Although MITgcm allows the automatic balance of the in/outflow through the boundaries we have decided to balance "offline" the barotropic velocity, simply computing the net flow through each boundary and then correcting the barotropic velocity profiles in each boundary in order to have a net inflow approximately zero. This has been done to avoid anomalous variation of sea level in the basin due to the unbalanced flow.

The average annual vertical profiles of the zonal component of current for the Western Boundary of the domain is shown in Figure 2.14 for both period 1987-1995 and 1996-2012.

\[10\] http://www.myocean.eu

31
Figure 2.13: Monthly Sea Surface Height simulated by MyOcean in October 1995 (a) and in October 2001 (b)
Figure 2.14: Average annual vertical profiles of the zonal component of current in $\frac{m}{s}$ applied to Western Boundary for the period 1987-1995 (a) and 1996-2012 (b)
Figure 2.14 shows that zonal flow in the area corresponding to the Sicily channel behaves as bi-layer with the inflow in the upper layer (in some points up to 200m) and an outflow in the rest of vertical profile. This imposed double layer behavior is clear observing also the Hovmoeller diagram of zonal component of current (Figure 2.15) for the period 1987-1995 (a) and 1996-2012 (b). This bilayer is present in both periods. It is worthwhile to mention that a computation of inflow through the Sicily channel has been done and has shown a mean value of 1 Sv for the period 1987-1995 and 1.1 Sv 1996-2012 which both fit with the known estimation of the inflow for this area [4]. Still the inflow appears to be stronger in Winter, Spring and Fall.

Figure 2.16 shows that zonal flow in the area corresponding to Cretan Passage and Kythira Strait. The Cretan passage shows an inflow along the Cretan coastlines and an outflow along the African coastlines for both periods of the analysis. In the case of the Kythira strait in the period 1987-1995 an inflow is observed along the northern flank of the strait while an outflow is observed along the southern flank. The situation appears in part to be reverted in the period 1996-2012.

The Hovmoeller diagram (Figure 2.17) for the Cretan Passage shows the upper layer characterized by an outflow corresponding to AW modified and inflow in the middle and deep layers in the period 1987-1995. For the period 1996-2012 the outflow occupies the first 1000m but this probably corresponds to the reinforce of outflow which is clear along the Africa coastline (Figure 2.16b). For the Kythira Strait (Figure 2.18) the inflow is prevalent in both periods but much stronger in the period 1996-2012 (Figure 2.18, b)
Figure 2.15: Hovmoeller diagram of zonal component of current in ($\text{m s}^{-1}$) in the Western Boundary corresponding to Sicily channel in the period 1987-1995 (a) and 1996-2012 (b)
Figure 2.16: Average annual vertical profiles of the zonal component of current in $\frac{m}{s}$ applied to Eastern Boundary for the period 1987-1995 (a) and 1996-2012(b)
Figure 2.17: Hovmoeller diagram of zonal component of current in ($\frac{m}{s}$) in the Eastern Boundary corresponding to Cretan Passage in the period 1987-1995 (a) and 1996-2012 (b)
Figure 2.18: Hovmoeller diagram of zonal component of current in (m s$^{-1}$) in the Eastern Boundary corresponding to the entrance of Kythira Strait in the period 1987-1995 (a) and 1996-2012 (b)
Experimental Boundary Conditions

The Temperature and Salinity data which have been employed as Initial Conditions and Boundary Conditions came from different datasets which cover different periods. The two periods which have been considered in our experiments are EMT-influence (1987-1995) \[51\] \[22\] \[27\] \[29\] \[50\] \[48\] \[59\] and EMT-relaxation/EMT free influence (1996-2012) \[59\] \[23\].

For the Boundary Conditions of Western and Eastern Open Boundary for the period of EMT-influence and for the Initial Conditions of the numerical experiments temperature and salinity data have been obtained from MEDAR/MEDATLAS II dataset. The objective of the project Medar/MedAtlas II is to make available a comprehensive data set of oceanographic parameters collected in the Mediterranean and Black Sea, through a wide cooperation of the Mediterranean and Black Sea countries increasing the exchange of data and informations between projects and the quality of data themselves. Vertical seasonal profiles for temperature and salinity are available for different areas of Mediterranean Area (Figure 2.19). The data consist mainly of vertical profiles divided by season.

Figure 2.19: Spatial coverage of Seasonal Vertical Profiles of Temperature and Salinity of Medar-Medatlas II dataset [http://doga.ogs.trieste.it/medar/climatologies/medz.html](http://doga.ogs.trieste.it/medar/climatologies/medz.html)

Figure 2.20 shows the vertical season profiles of Temperature (a) and Salinity (b) (interpolated to fit our vertical resolution) for the Sicily channel (DI3, Figure 2.19). Assuming that a possible tendency in temperature and salinity for this area is negligible over decadal scale these data have been applied to the Western Boundary as a "perpetual" year with always the same four profiles, one for each season and for both periods EMT-influence and EMT-relaxation/EMT free influence.
Figure 2.20: Seasonal vertical profiles of Temperature (in C) (a) and Salinity (in psu) (b) in the Sicily Channel according to Medar-Medatlas II dataset in winter (red), spring (blue), summer(green), fall (fuchsia)
In the case of the Eastern Boundary (Figure 2.21, Figure 2.22) a different approach has been followed: due a small coverage of data the profiles exhibited (mainly in the case of salinity, Figure 2.22) “noisy” behavior (red lines). Thus we decided to consider an average profile for both Cretan Passage (DH3, Figure 2.19) and Kythira Strait (DH2, Figure 2.19) given by the median of data in each depth of our profiles (blue lines). These median profiles have been applied perpetually to the Eastern Boundary in the period of EMT-influence.
Figure 2.21: Seasonal vertical profiles (in red) and median profile (blue) of Temperature (in °C) in the Kythira Strait (a) and Cretan Passage (b) according to Medar-Medatlas II dataset.
Figure 2.22: Seasonal vertical profiles (in red) and median profile (blue) of Salinity (in psu) in the Kythira Strait (a) and Cretan Passage (b) according to Medar-Medatlas II dataset
For the period EMT-relaxation/EMT free influence the most recent data available as Boundary Conditions for Temperature and Salinity derive from the ARGO-float dataset. Argo is a global array of more than 3,000 free-drifting profiling floats that measures temperature and salinity of the upper 2000 m of the ocean. Figure 2.23(a) shows the distribution of float in the Mediterranean Sea updated at 22 January 2015. Figure 2.23(b) shows their distribution in the area of the Eastern Boundary for the period 2005-2013.

In this analysis a similar approach to that followed for Medar-Medatlas dataset in the Eastern Boundary is followed. Figure 2.24a,b and Figure 2.25a,b show respectively the vertical profiles (red line) of Temperature and Salinity in the Cretan Passage and Kythira Strait and their respectively median (blue line). The green line are additional data available in the area and coming from the NODC center hosted at OGS. These median profiles have been applied as boundary conditions at the Kythira strait and Cretan Passage as perpetual year for the period EMT-relaxation/EMT free influence.

\footnote{http://www.argo.ucsd.edu}
\footnote{http://nodc.ogs.trieste.it/nodc/}
Figure 2.23: Spatial distribution of the float in the Mediterranean region updated at 22 January 2015 (from \url{http://nettuno.ogs.trieste.it/sire/medargo/active/index.php}) (a) and in the area of Eastern Boundary in the period 2005-2013 (b).
Figure 2.24: Vertical profiles of Temperature (in C) in the Kythira Strait (a) and Cretan Passage (b) according to Float-Argo.
Figure 2.25: Vertical profiles of Salinity (in psu) in the Kythira Strait (a) and Cretan Passage (b) according to Float-Argo
Finally another class of Temperature and Salinity data has been considered in this work despite it has not been employed as boundary conditions. These data have been kindly provided by prof. Wolfgang Roether from the University of Bremen and have been taken during four German cruises which took place in the Mediterranean in the 1987, 1995, 1999 and 2011 ([50], Table 2.2).

Figure 2.26 shows the vertical temperature and salinity profiles in the Cretan Passage (a,b) and in the Kythira Strait (c,d) : it is clear the signal of EMT in the Cretan passage and Kythira deep waters with an increase of temperature and salinity between 1987 and 1995 [61] [51] [22] [27] [29] [50] [48].

<table>
<thead>
<tr>
<th>Name of Cruise</th>
<th>year</th>
<th>period of the year</th>
</tr>
</thead>
<tbody>
<tr>
<td>M5/6</td>
<td>1987</td>
<td>August-September</td>
</tr>
<tr>
<td>M31/1</td>
<td>1995</td>
<td>January-February</td>
</tr>
<tr>
<td>M44/4</td>
<td>1999</td>
<td>April-May</td>
</tr>
<tr>
<td>M51/2</td>
<td>2011</td>
<td>October-November</td>
</tr>
</tbody>
</table>

Table 2.2: Name, year and period of the German cruises
Figure 2.26: Vertical profiles of Salinity (in psu) and Temperature (in °C) in the Cretan Passage (a,b) and Kythira Strait (c,d) according to the data provided by W. Roether.
Additional Boundary Conditions

Surface heat forcing has been chosen in order to reproduce the DWF in the SA in winter and to guarantee the energy balance of the system in the most advanced experiments (OPWIND). A flux of $-800 \text{ W m}^{-2}$ has been set for two months (January-February) over the SA in the case of CLNOWIND and for one month (January) in the case of OPWIND. The choice for such strong heat flux derives from the necessity to compensate the absence of Northern Adriatic Deep Water and to accelerate the DWF in our experiments. Still in the case of OPWIND experiments this loss of heat is balanced with an uniform flux over the domain of $\sim 1.61 \text{ W m}^{-2}$. Figure 2.27 shows a comparison of meridional transport through the Otranto Strait for CLWIND and OPWIND experiments: in the first it is noticeable that after the spin up the most realistic values of average meridional transport concern OPWIND with values of $\sim 0.2 \text{ Sv}$. Higher values of meridional transport in the CLNOWIND experiments are not only a consequence of stronger heat forcing but also because this flux is not balanced. Not balancing the heat flux in winter means increasing the density of the surface waters so enhancing the potential for DWF over SA which led in some cases (as in the case of CLNOWIND experiments after ten years, not shown in Figure 2.27) to great value of outflow at Otranto Strait, more than 1 Sv in some cases. These values have been observed only with extremely harsh winter as in 2012. No precipitation, evaporation and run off fluxes have been added as in the literature there are not available scientific papers about the importance of these forcing in the Ionian Sea.

Still as additional boundary conditions no slip conditions for the momentum have been im-

---

\[14\text{the flux is zero for the rest of the year}\]
Table 2.3: Values of diffusivity coefficient for momentum, temperature and salinity

<table>
<thead>
<tr>
<th>ViscAh</th>
<th>$400 , \text{m}^2/\text{s}$ (CLNOWIND, Idealized bathymetry) / $100 , \text{m}^2/\text{s}$ real bathymetry</th>
</tr>
</thead>
<tbody>
<tr>
<td>ViscAz</td>
<td>$10^{-4} , \text{m}^2/\text{s}$ (CLNOWIND, Idealized bathymetry) / $10^{-3} , \text{m}^2/\text{s}$ real bathymetry</td>
</tr>
<tr>
<td>DiffkhT</td>
<td>$400 , \text{m}^2/\text{s}$</td>
</tr>
<tr>
<td>DiffkzT</td>
<td>$10^{-4} , \text{m}^2/\text{s}$</td>
</tr>
<tr>
<td>DiffkhS</td>
<td>$400 , \text{m}^2/\text{s}$</td>
</tr>
<tr>
<td>DiffkzS</td>
<td>$10^{-4} , \text{m}^2/\text{s}$</td>
</tr>
</tbody>
</table>

Experimental Initial Conditions

Initial Temperature and Salinity conditions (Figure 2.28) came again from the Medar-MedAtlas II dataset and are obtained extending horizontally and uniformly over the basin the median of vertical seasonal profiles for the areas DJ3-DJ4-DJ5-DJ6-DJ7-DJ8 (Figure 2.19). Initial velocity is null all over the domain. Numerical stability of the experiments, even in presence of sharp initial temperature/salinity/velocity gradients between the domain and its open boundaries has been guaranteed by the sponge layers (see previous paragraph for more details).
Figure 2.28: Initial Conditions: vertical profile of Temperature (in °C) (a) and Salinity (in psu) (b)
Chapter 3

Results

This chapter provides in form of draft paper the results of the analysis of the outputs of the numerical experiments
A coarse resolution primitive equation numerical model based on the MIT general circulation model (MITgcm), is used to study the baroclinic dynamics of the Adriatic-Ionian System (AIS) in order to assess the relative importance of remote forcings (wind stress and thermohaline fluxes) on the vorticity and energy budget of the system. An approach based on an increasing complexity in the model forcings and domain has resulted in a more extensive understanding of AIS dynamics. It is shown that the influence of the Adriatic Sea outflow on the Ionian Sea dynamics is not affected by the bathymetry of the basin, resulting in both cases in a bi-layer structure for vorticity in the Ionian Sea induced by vortex stretching mechanism. Vorticity and energy balance of AIS appears to be more influenced by the Levantine Basin: this influence is strictly correlated with substantial inflow of Available Potential Energy (APE$	ext{QG}$) in AIS from the Cretan Passage / Kythira Strait. The wind itself appears to have a marginal role in the vorticity/energy budget of the system: it is able to reinforce/weaken the circulation but not to induce changes in sign in the circulation. The characteristics of this inflow strongly depend on the thermohaline properties of waters entering in the region from east. Significant changes in the potential density profiles of these waters mainly at level of surface/intermediate layers (up to 400m) have been observed during the reversal of circulation which took place in the Ionian in 1997, 2006 and 2011.

I. Introduction

The Adriatic-Ionian System (hereafter AIS, fig.1) is formed by the Adriatic Sea and the Ionian Sea and plays an important role in the Mediterranean Thermohaline circulation (MTHC) component in the Eastern Mediterranean Sea (EMED). The Adriatic [18] [13] is an elongated basin (approximately 800 km long) which is characterized by a shallow northern part (NA, Northern Adriatic, approximately 100 m deep) and southern deep part (SAP, Southern Adriatic Pit approximately 1200m). It is connected through the Otranto Strait (75km wide and approximately 800m deep) with the Ionian Sea.

Both shallow northern part (Northern Adriatic, NA) and deep southern part (Southern Adriatic, SA) are interested by deep waters formation processes (DWP) which are favored by the cold and dry wind Bora flowing over the basin mainly in winter and the presence of Levantine Intermediate Water (LIW) which increase the potential for deep waters forma-
tion [18] [13] [19]. A major part of deep waters (about 82%) are produced in the SA [18] [19] through an open convection process [20] and are mixed with those produced in the NA, finally flowing through the Otranto Strait with an outflow ranging between 0.1 Sv and 0.4 Sv [20], oxygenating the deep layers of Ionian Sea and the EMED. Usually the so-formed Adriatic Deep Waters (ADW) are characterized by a temperature of $T < 13.3 \text{C} \,$, a salinity $S > 38.68$ psu and $\sigma_\theta \sim 29.17 - 29.18 \text{ kg m}^3$ [8].

The Ionian Sea is the deepest basin of Mediterranean Sea. It is connected through the Sicily channel with the Western Mediterranean, with the Adriatic Sea through the Otranto strait, with the Cretan Sea through Kythira/Antikythira Straits and with the Levantine Basin through the Cretan Passage. It represents a sort of crossroad for the main kind of waters being part of MTHC [7]. Its deep layers are filled with the ADW, its intermediate layers are filled with LIW while its surface layers are influenced by relatively fresher and colder AW moving eastwards. A change of the MTHC component in EMED occurred during the 1990s when the main source of deep waters for the basin shifted from the Adriatic to the Aegean Sea [27] [28][14] [16]. This event has been called Eastern Mediterranean Transient (hereafter EMT) and it has been attributed to meteorological anomalous conditions in the area and to changes in the circulation patterns [27] [28][14] [16]. During EMT the deep layers of EMED were filled with an inflow of 3 Sv (greater with respect to the Adriatic inflow) [28] for seven years with waters with greater temperatures and salinity values ($T > 13.7\text{C}, S > 38.8 \text{ psu and } \sigma_\theta > 29.18 \frac{\text{kg}}{\text{m}^3}$) [14] with respect to those produced in the Adriatic Sea. After 2000 the DWF shifted back to the Adriatic despite been recently shown that thermohaline properties in the Ionian are still far from the observed pre-EMT phase [4].

The stationarity of the thermohaline circulation in the EMED was argued with the proposed mechanisms of BiOS (Bimodal Adriatic-Ionian Oscillating System) [7]. This mechanism has been suggested to explain the variability observed in the sign of NIG (Northern Ionian Gyre) which on decadal scale reverses from anticyclonic to cyclonic and vice versa. This decadal variability, affecting the spreading of the AW eastwards and LIW westwards, is able to influence the salinity in the Eastern Basin [9], in the Sicily channel [11] and the advection of nutrients and salt in the Adriatic Sea [6] [7]. The variability observed has been often associated with changes in the wind stress curl [15] [25] [23]. On the other hand [7] this variability has been explained in term of BiOS (Bimodal Adriatic-Ionian Oscillating System). BiOS is a feedback mechanism linking the Ionian and Adriatic according to:

- AIS behaves as a bimodal oscillating system
- when anticyclonic circulation are present in Ionian, AW are deflected in the Adriatic leading to production of ADW of lowering density which spread in the Ionian producing a deepening of isopycnal surfaces in the Ionian and stretching of water column resulting in a reversion from anticyclonic to cyclonic conditions.
- when cyclonic circulation are present in Ionian salty LIW enters in the Adriatic leading the production of ADW of increasing density which produce a shallowing of isopycnal surfaces in the Ionian and so a reversion from cyclonic to anticyclonic conditions.

This mechanism suggests that internal processes are able to sustain the variability of the circulation in the Ionian independently from external forcings such as wind [7]. The Adriatic Sea acts a source of potential vorticity for the Ionian [30] while Ionian is characterized by an approximate bi-layer structure with the two layers characterized by opposite sign in the vorticity [7]. This two layers structure is not influenced by the topography and the zero-level of vorticity is located under 1000 m [7]. Altimetric maps shows an anticyclonic mode starting in 2006 [7], a cyclonic mode starting in 2011 and
unexpectedly ending in 2012 when a new reversal to anticyclonic mode has been observed [10]. This premature inversion has been attributed to extremely strong winter in 2012, which caused the formation of very dense Adriatic waters, flooding Ionian flanks in May and inverting the bottom pressure gradient [10].

The aim of this study is to provide a deeper insight into the dynamics of the AIS through a modeling approach in order to assess the relative impact of the external forcing/internal processes acting on AIS by evaluating their effects on vorticity and energy budget. This work is organized as follows: section II provides the description of numerical model used and the forcings used in the numerical experiments, section III provides an analysis of the vorticity and energy budget of the AIS; finally section IV provides the final conclusions of the work.

II. Model and methods

I. Numerical Model (MITgcm)

The dynamics of AIS has been simulated through a model based on primitive equations, MITgcm (MIT general circulation model) [1]). MITgcm is a state of the art three-dimensional model [1] that can be used to simulate, on a wide range of computational platforms, the dynamics of the atmosphere and the ocean over a wide range of spatial and temporal scales. It employs a finite volume approach to discretize the momentum equations and the partial cells to treat complex geometries.

Assuming an incompressible flow and the Boussinesq approximation the equations solved by MITgcm are [21][22]:

- Momentum:
  \[ \rho_0 D_t \vec{u} + 2\Omega \wedge \rho_0 \vec{u} + g\rho_0 k + \nabla p = F \]  
  (1)

- Continuity:
  \[ \rho_0 \nabla \cdot \vec{u} = 0 \]  
  (2)

- Mass Conservation:
  \[ \partial_t \eta + \nabla \cdot (H + \eta) \bar{u}_h = \text{Prec} - \text{Evap} + \text{Runoff} \]  
  (3)

- Energy Conservation:
  \[ D_t T = A_T \nabla^2 T + Q_T \]  
  (4)

- Salt Conservation:
  \[ D_t S = A_S \nabla^2 S + Q_S \]  
  (5)

- State equation:
  \[ \rho = \rho(S, p, T) \]  
  (6)

where \( \vec{u} \) is the three dimensional velocity vector, \( p \) is pressure, \( \rho \) is in-situ density, \( \eta \) is the displacement of the free-surface from the sea level at the rest, \( T \) is the potential temperature and \( S \) is salinity, \( \rho_0 \) is a constant reference density, \( g \) is the constant gravitational acceleration and \( H \) is a fixed-in-time bottom depth [21] [22]. \( F \), \( \text{Prec-Evap+Runoff} \), \( Q_T \), and \( Q_S \) are the forcing fields. \( \nabla^2 T \) and \( \nabla^2 S \) are the diffusive fluxes of heat and salt, \( A_T \) and \( A_S \) are the horizontal eddy diffusive coefficients for the temperature and salinity. \( D \) is the material operator (i.e : \( \frac{D}{Dt} + \bar{u} \cdot \nabla \)). MITgcm is able to make the hydrostatic approximation to the vertical momentum equation [21] [22]. The equations are integrated in the space over space-filling finite volumes (cells) and in time according to a quasi-second order Adams-Bashfort scheme. The resulting horizontal velocities components are staggered on a Arakawa C-grid while the vertical component on a Lorentz grid.

II. Model configuration and Boundary conditions of the model

II.1 Experimental protocol

A series of numerical experiments have been set with the following protocol:

- to set up an idealized model for the AIS
- to progressively refine the physics and the forcings of the model in order to rank the relevant mechanisms acting on the variability observed in the Ionian circulation
• to define which is the contribution of the remote forcing (Wind) to the Ionian circulation
• to assess the Vorticity and Energy Balance of the AIS (and the Ionian in particular)

II.2 Model configuration

The model has been run with an Hydrostatic configuration. Both available parametrization for the upper surface boundary, i.e Rigid Lid and Linear Free Surface have been used. The grid chosen for the numerical experiments is Spherical Polar. For the advection of tracer the scheme 33 has been employed in order to guarantee the numerical stability of runs and to avoid anomalous advection of tracers [Querin S., personal communication]. This scheme is essentially a 3\textsuperscript{rd} order DST (direct space time) scheme with flux limiter.

II.3 Turbulence closure

The model used a constant both horizontal/vertical eddy diffusivity for momentum, temperature and salinity. The choice of constant vertical eddy diffusivity parameters has been a consequence of problems experimented with the numerical stability of the model but also of the assumptions and parametrization employed in the first experiments. For consistency the values have been maintained constant throughout all the experiments despite the most advanced experiments use as forcing the wind. The values of these coefficients will be given in the section III.

II.4 Datasets

We adopted as boundary conditions two main kind of data: Model and Experimental data.

Model Boundary Conditions Zonal and meridional component of wind speed which we have prescribed as boundary conditions at the surface of the domain come from RegCM4 simulations carried out in the Med-Cordex domain [12] [29]. The RegCM4 (The Regional Climate Model system v4), currently developed and maintained in the Earth System Physics (ESP) section of the ICTP [12], is a flexible, portable and easy to use Regional climate model which can be used in any region of the World and in a wide range of studies with grid spacing until 10 km [12]. The monthly data with a resolution of 50km are available for the period 1987-2012. The area corresponding approximately to AIS has been extracted from the dataset and then interpolated over a regular longitude-latitude grid. As it has been shown in [3] [7] a mean anticyclonic input from wind stress characterized the Ionian area after 1994. Because of this a climatology (12 months) of zonal and meridional component of wind speed has been computed for the period 1987-1995 and 1996-2012.\footnote{The reasons for the choices of these two periods will be discussed in the next section.} Fig.2 shows the mean wind stress curl divided by $\rho$ and $H=500$m over the AIS in the period 1987-1995 (a) and 1996-2012 (b).
Zonal and meridional component of current which we have prescribed at boundaries of the domain have been derived from MyOcean dataset. MyOcean is an extended dataset coming from the simulations of Mediterranean Forecasting System based on the NEMO model. Nemo has been implemented on the Mediterranean Domain with a horizontal spatial resolution of $1/16^0 \times 1/16^0$ and 72 unevenly spaced vertical levels [24]. The data extracted from the dataset has a monthly resolution and have been interpolated along the horizontal direction and along the vertical direction in order to fit our spatial horizontal/vertical resolution. As in the case of wind speed component a climatology (12 months) of zonal and meridional component has been computed for the period 1987-1995 and 1996-2012. Both climatologies have been corrected in order to balance the inflow/outflow through the boundaries (despite this option is available in the MITgcm) in order to avoid unrealistic variation in the level of the basin.

**Experimental Boundary and Initial Conditions** Temperature and Salinity vertical profiles, prescribed as boundary conditions at two boundaries, have been derived for the period 1987-1995 from Medar/Medatlas II dataset [17] and for the period 1996-2012 MedArgo Float project dataset [31]. The two periods considered in our experiments take into the EMT (1987-1995) phase in MTHC [7] [14] [16][27] [28] and the EMT-relaxation/EMT free influence phase (1996-2012) [28] [4]. Medar-Medatlas II is an international project which aims to make available wide dataset of oceanographic parameters collected in the Mediterranean and Black Sea through an improvement of the cooperation of the Mediterranean and Black Sea countries. The data are available as seasonal/annual profiles: for the Sicily Channel all the four seasonal profiles have been considered and repeated as perpetual year for the entire period 1987-2012 (fig. 3,a,b). For the period 1987-1995 for Cretan Sea and Cretan Passage the median of the four seasonal profiles has been taken into account (fig.4, red line) and applied as perpetual year. Due to the fact that the AIS communicates through the Kythira Strait with the Cretan Sea, from now on, the latter will be identified as Kythira Strait. For the period 1996-2012, as in the case of Medar/Medatlas II data, the median of the vertical profiles have been computed for both Kythira Strait and Cretan Passage (fig.4, blue line) and applied as perpetual year.
Figure 3: Seasonal vertical profiles of Temperature (a, in °C) and Salinity (b, in psu) in the Sicily Channel from Medar-MedAtlas in winter (red line), spring (blue line), summer (green line) and fall (fuchsia line).
Figure 4: Vertical profiles of Temperature (a, b in C) and Salinity (c,d in psu) in the Kythira Strait (a,c) and Cretan Passage (b,d) according to Medar (red line) and float data (blue line).

Initial Temperature and Salinity conditions (the same for all the experiments) are obtained from the Medar-Medatlas II dataset: the median of vertical profiles available for the AIS has been computed and applied all over the domain of the integration. Initial velocity is zero all over the domain.

II.5 Additional Boundary Conditions

As additional surface boundary conditions an artificial heat flux has been imposed in order to reproduce the DWF in the SA: the values of this heat flux was varied in order to guarantee the realistic value of outflow at Otranto Strait and to compensate the absence of DWF in the NA. More details about these heat fluxes
are provided in section III. Together with prescribed zonal and meridional component of speed current a sponge layer algorithm have been applied to both boundaries in order to ensure the numerical stability: in the sponge layer (10 grid points width) both velocities components in each cell of each vertical level are an weighted average between the velocity of the innermost point of the sponge layer and the velocity of the corresponding open boundary cell with the weight defined by the distance from the boundary. The velocities in the inner grid points are relaxed to these prescribed values in a 5 days\(^{-1}\) time period. Additional boundary conditions (for all the experiments) are no slip conditions for momentum at the solid boundaries while due to coarse resolution of the experiment, free slip conditions on the bottom is imposed.

II.6 Definition of Vorticity and Energy Budget in the basin

**Vorticity Balance** Assuming the hydrostatic approximation holds for a two dimensional flows the momentum equation will be:

\[
pD_t \tilde{u}_h + 2\Omega \wedge \tilde{u}_h + \tau \nabla p + \nabla \cdot \mathbf{v} = \frac{\eta}{H} + \mu \nabla^2 \tilde{u}_h
\]

where \(\tau\) is wind stress in \(N/\text{m}^2\) and \(\mu\) is the dynamical viscosity in \(m^2/\text{s}\) and \(\nabla^2 \tilde{u}_h\) the laplacian of horizontal velocity field and \(H\) the thickness of the layer of the fluid (for example 500 m). Applying the curl operator \((\nabla \wedge)\) to (7) we found:

\[
\partial_t \zeta = (\zeta \cdot \nabla) \tilde{u}_h - \zeta \nabla \cdot \tilde{u}_h - (\tilde{u}_h \cdot \nabla) \zeta + \frac{\nabla \cdot \nabla \mathbf{v}}{\rho} + \frac{\nabla \cdot \nabla p}{\rho \mu} + \nu \nabla^2 \zeta
\]  

(8)

The equation (8) is the well known Vorticity Balance equation: where \(\zeta\) is the vertical component of vorticity, \((\zeta \cdot \nabla) \tilde{u}_h\) is the tilting term\(^2\), \(\zeta \nabla \cdot \tilde{u}_h\) is the stretching/squeezing term, \((\tilde{u}_h \cdot \nabla) \zeta\) the advection term, \(\frac{\nabla \cdot \nabla \mathbf{v}}{\rho}\) the baroclinicity term, \(\frac{\nabla \cdot \nabla p}{\rho \mu}\) the input of vorticity from the wind stress and finally \(\nu \nabla^2 \zeta\) is the diffusion of vorticity. \(\zeta\) is available as output from the MITgcm: the others terms of (8) can be estimated using the diagnostics of momentum balance provided by the model. In fact the different horizontal momentum terms are split into different diagnostics, mainly to reflect the time-stepping options (forward in time, backward in time, Adams-Bashforth) [JM Campin, personal communication]. The numerical momentum balance for the \(u^*\) (numerical zonal component of speed) using the MITgcm diagnostics can be written [JM Campin, personal communication] as:

\[
\partial_t u^* = -g \delta \eta^* \frac{\partial \eta}{\partial x} + u^* \frac{\partial u^*}{\partial x} + u^* \frac{\partial u^*}{\partial y} + \frac{u^*}{H} \nabla^2 \eta^* + u^* \mid_{\text{external forcings}} + u^*_{\text{AB}}
\]  

(9)

Equation (9) shows that the total numerical momentum tendency for \(u^*\) is given by the sum of the numerical momentum tendency from horizontal surface pressure gradient (Rigid Lid/Free Surface), from the horizontal hydrostatic pressure gradient, from Advection, Coriolis, Viscous Diffusion, External Forcing and Time Stepping (AB, Adams-Bashforth scheme, despite this term is negligible with time averaging over a long period). \(\eta^*\) is the numerical sea surface height anomaly and \(g\) is the gravity acceleration. The same numerical balance holds for the \(v^*\) (numerical meridional component of speed). In order to find the corresponding numerical vorticity balance equation we apply the curl operator discretized as in the MITgcm. The numerical discretization of curl for the horizontal velocity field is given as:

\[
\frac{1}{A_c} (\delta \Delta y_i \delta v^* - \delta \Delta x_i u^*)
\]

(10)

- \(A_c\) area of vorticity cell
- \(\delta \Delta y_i \delta v^*\) equal to \(v^*_i - v^*_{i-1} \delta \Delta y_{c_{i-1}}\)
- \(\delta \Delta x_i u^*\) equal to \(u^*_i - u^*_{i-1} \delta \Delta x_{c_{i-1}}\)

with \(\Delta y_c\) and \(\Delta x_c\) the distance between the centers of cell along x and y direction respectively. The terms obtained can be multiplied for the corresponding frequency of snapshots \(2\) this term has been found to be always negligible through a dimensional analysis (not shown here).
of the model (for example 432000s) in order to get the input of vorticity (in $s^{-1}$). The resulting vorticity input $\zeta_{\text{input}}$ can be integrated over a volume established (taking into account the spherical coordinates) for the domain (for example down to 500m) in order to get an average vorticity for unit of volume ($\zeta_{\text{input}}$, in $s^{-1}$) defined as:

$$
\frac{1}{\Delta V_{\text{domain}}} \int \int \int \zeta_{\text{input}} \partial r \partial \phi \partial \theta
$$

Figure 5: Reconstructed (in $s^{-1}$) $\zeta$ (red line) vs $\zeta$ computed by the model (blue line) in the first 500m of the Ionian Sea

This approach provides a proper estimation of the stretching, diffusive, and advection vorticity terms but some troubles have been encountered with the estimation of baroclinicity term. In fact vertical/horizontal spatial resolutions are too coarse in order to properly estimate the baroclinic term and, instead, the cross product between the two gradients vector should require a very fine vertical resolution of the model in order to provide a proper estimation of the resulting vector [Falcini F. and G.Sannino, personal communication]. The cumulated input of vorticity from baroclinicity term has been taken as difference between $\zeta$ provided by model and the cumulated input of vorticity from the other terms of the budget.

Fig. 5 shows the comparison between the reconstructed $\zeta$ and the output provided by the model during a simulation which was run for 500 days forced only by heat flux: an analysis of vorticity balance has shown that stretching/squeezing (average contribution $-7.29 \cdot 10^{-9}s^{-1}$) and the diffusivity term (average contribution $6.97 \cdot 10^{-9}s^{-1}$) are prevalent with respect to the other terms to determine the behavior of $\zeta$.

Energy Budget A simple approach which could be used to describe the energy budget of a basin in response to external forcings/internal processes is to consider a two components energy cycle [31]:

$$
\frac{dKE}{dt} = C(APE_{QG}, KE) + G_{KE} - D_{KE}
$$

$$
\frac{dAPE_{QG}}{dt} = -C(APE_{QG}, KE) + G_{APE_{QG}} - D_{APE_{QG}}
$$

where KE is the kinetic energy (in J), $APE_{QG}$ (in J) is the quasi-geostrophic available potential energy, $C(APE_{QG}, K)$ is the conversion term (in W) between $APE_{QG}$ and KE, $G_{KE}$ (in W) is the mechanical power input by the wind, $D_{K}$ is the viscous dissipation (in W), $G_{APE_{QG}}$ (in W) is the $APE_{QG}$ generation by surface buoyancy fluxes while $D_{A}$ is its dissipation (in W) due to turbulent molecular diffusive fluxes of heat and salt [31]. $APE_{QG}$ is that part of the potential energy that can be potentially transformed into kinetic energy through mechanisms concerning either the barotropic or baroclinic instability [15]. From mathematical point of view there are several ways to compute the $APE_{QG}$ in a domain. In this work we use the approach of [2]:

$$
-8 \int \int \int \frac{|\rho - \bar{\rho}(z)|^2}{2\rho_z} \partial r \partial \phi \partial \theta
$$

where $\rho$ is the local potential density, $\bar{\rho}(z)$ the average potential density in that level, $\rho_z$ is the vertical gradient of potential density. $G_{KE}$ [2] is equal to:

$$
\int \int \tilde{\tau} \tilde{u}_h \partial \phi \partial \theta
$$

where $\tilde{\tau}$ is the wind stress and $\tilde{u}_h$ is the horizontal velocity fields.$C(APE_{QG}, KE)$ is given
Figure 6: Domain of integration for the IdeA(a) and ReaL (b) experiments in long-latitude coordinates

by [31]:

\[- \int \int \int \vec{u}_h \nabla h \partial r \partial \phi \partial \theta\]  

\(G_{APEQG}\) is given by \(^3\):

\[\int \int \int \frac{\alpha}{C_p} Q_h \partial r \partial \phi \partial \theta\]  

III. RESULTS

I. Effects of DWF and topography on the vorticity of the Ionian

In this section we discuss the effects of DWF (in absence of other kind of external forcings such as wind or later fluxes ) on the vorticity budget of the Ionian Sea.In order to check whether these budget may be affected by topographic effects two classes of experiments have been prepared : one with a ideal coarse bathymetry (IdeA) and one with real higher resolution bathymetry (ReaL).

\(^3\)In order to get uniform measure units among all the quantities the original definition [5] has been multiplied for \(\rho_o\) and integrated over a thickness of fluid corresponding to the first level in the model

where \(C_p\) is specific heat capacity of water and is equal to 3990 J\(K^{-1}C^{-1}\), \(\alpha\) is the coefficient of thermal expansion equal to \(2.4 \times 10^{-4} C^{-1}\). The sign of \(Q\) is positive when heat is entering the ocean . The remain terms \(D_{KE}\) and \(D_{APEQG}\) are taken as differences between the rest of the terms in the equations (12) (13) but they won’t be considered here.

I.1 Domain of integration

Fig.6 shows the domain of integration for this two classes of numerical experiments. IdeA (a) is characterized by a Ionian with a flat bottom with a maximum depth of 2000m (blue rectangular) and connected with the SA (yellow square) through an ipotetical and flat Otranto Strait. NA and Middle Adriatic are excluded from our computational domain (also in the case of ReaL).The number of points is \(17 \cdot 49\) for 24 levels , \(0.25^o\) is its horizontal spatial resolution. The number of levels is higher in the first 200m (level spacing 20 m-50 m) and then decreases with the depth ( 200 m in the last level). ReaL (b) is a real topography coming from the Mediterranean Forecast System ( par. II.4 , [24]). The bathymetry is complex
in this case and the maximum depth for the domain is 4000 m. The domain is extended over 96 × 96 points (0.125° in long-lat) with 32 levels spaced as in the case of IdeaA but with the last level spaced of 250 m.

I.2 Model Configuration and Boundary Conditions

The Model Configuration is the same in both cases as reported in II.2 and use constant horizontal/vertical diffusivity coefficients whose values are reported in Tab. 1. The model is always forced at the surface essentially by an artificial heat flux not balanced all along the simulations. This flux has a value of -800 W m⁻² over SA (and zero all over the rest of the domain) and acts on this region in January and February. It is zero for the rest of the year. This high value has been chosen in order to accelerate the DWF in the region but also to compensate the absence of NA in our computational domain. The lack of balancing in the heat flux has led to a progressive increase of potential in DWF due to the continuous increase of density. This has resulted in an outflow at Otranto strait after 10 years of simulation greater than 1 Sv, value observed only in correspondence of extremely harsh winter as in 2012 [10]. The wind forcing is null all over the domain. We considered 5-days mean data in both cases.

Table 1: Diffusivity coefficients (h=horizontal, z=vertical) for Momentum, Temperature and Salinity in the IdeA and ReaL case

<table>
<thead>
<tr>
<th>Parameter</th>
<th>IdeA</th>
<th>ReaL</th>
</tr>
</thead>
<tbody>
<tr>
<td>ViscAh</td>
<td>400 m² s⁻¹</td>
<td>100 m² s⁻¹</td>
</tr>
<tr>
<td>ViscAz</td>
<td>10⁻⁴ m² s⁻¹</td>
<td>10⁻³ m² s⁻¹</td>
</tr>
<tr>
<td>DiffkhT</td>
<td>400 m² s⁻¹</td>
<td>400 m² s⁻¹</td>
</tr>
<tr>
<td>DiffkzT</td>
<td>10⁻⁴ m² s⁻¹</td>
<td>10⁻⁴ m² s⁻¹</td>
</tr>
<tr>
<td>DiffkhS</td>
<td>400 m² s⁻¹</td>
<td>400 m² s⁻¹</td>
</tr>
<tr>
<td>DiffkzS</td>
<td>10⁻⁴ m² s⁻¹</td>
<td>10⁻⁴ m² s⁻¹</td>
</tr>
</tbody>
</table>

I.3 Effects of DWF and topography on the vorticity of the Ionian Sea

Fig. 7 a shows \( \zeta \) in the first 500 m in the Ionian Sea between 16E-19.5E and 33.25N-38N for the IdeA case (blue line) and ReaL case (red line). The mean value of \( \zeta \) is equal to \(-3.33 \times 10^{-8} \text{s}^{-1}\) and \(-3.93 \times 10^{-8} \text{s}^{-1}\) respectively. No changes in the signs of vorticity in the area have been observed. The DWF process in the SA induce always an anticyclonic circulation in the Ionian Sea: this is clear from Fig. 7 b,c where the Hovmoeller diagram of \( \zeta \) for the Idea (a) and ReaL (b) experiments are shown. In both cases the Ionian Sea behaves as a bi-layer with opposite sign in vorticity. The upper layer (up to 1200-1500m) is characterized by a permanent negative sign in \( \zeta \) corresponding to a clockwise circulation. Under 1500m the sign is permanent positive (counterclockwise). The zero-level for vorticity is located under 1000m (as predicted by [7]).
The bi-layer structure can be easily visualized through a passive tracer released in the SA. In the first 500 m (Fig. 8) a,b,c the temporal distribution follows a clockwise circulation. In the last 500 m (Fig. 8) d,e,f the tracer shows clearly that the Deep Waters formed in the SA follow a counterclockwise circulation filling the deep layers of the Ionian. These results support the idea that the Adriatic Sea is able to induce a clockwise circulation in the Ionian as a consequence of the production of high volumes of dense waters due to the particular harsh winter conditions as in 2012 [10]. Table 2 shows the main component (the others have been omitted as negligible) of vorticity budget of Ionian Sea in the first 500m. The values are the cumulated values for the entire experiment. It is clear the strong differences between the two classes of experiments produced by the variation in both horizontal and vertical momentum diffusion coefficients. In the IdeA experiment the diffusion is the strongest term: only a combined stretching/advection term leads to an general negative input of vorticity for the system. In the case of ReaL, the stretching term is the main term of the budget. The combined term diffusion/advection is not able to balance the stretching and so the general input of vorticity is always negative. All the other terms are negligible.

Table 2: Vorticity budget: cumulated input (in s\(^{-1}\)) in the IdeA and Real case

<table>
<thead>
<tr>
<th>Vorticity Budget Component</th>
<th>IdeA</th>
<th>Real</th>
</tr>
</thead>
<tbody>
<tr>
<td>Advection</td>
<td>(-2.30 \times 10^{-7})</td>
<td>(2.27 \times 10^{-6})</td>
</tr>
<tr>
<td>Diffusion</td>
<td>(5.44 \times 10^{-6})</td>
<td>(5.83 \times 10^{-7})</td>
</tr>
<tr>
<td>Stretching</td>
<td>(-5.23 \times 10^{-6})</td>
<td>(-2.88 \times 10^{-6})</td>
</tr>
</tbody>
</table>
Figure 8: Spatial distribution of a passive tracer between 20m and 500 m (a,b,c) and 3500 and 4000m (d,e,f) during the ReaL experiment.
II. Effects of the External forcings on the vorticity of the Ionian

In this section we discuss the effects of the external forcing such as AW, LIW and wind on the vorticity budget of the Ionian Sea.

II.1 Domain of integration

Fig. 9 shows the domain of integration for this class of numerical experiments. It is a real bathymetry derived, as in the case of ReaL, from the Mediterranean Forecast System (par. II.4, [24]). The domain is extended over 108 \( \times \) 100 points (0.125° in long-lat) with 32 levels spaced as in the case of ReaL. The domain of integration has two open boundaries. The first boundary is on the western side of the domain connecting Sicily with Northern Africa, corresponding to the Sicily Channel. The other boundary is on the Eastern side of the domain and is divided in two channels: one connecting the Ionian with the Cretan Sea through the Kythira Strait (hereafter, as stated before, Kythira Strait) and one connecting the Ionian with the Levantine Basin (Cretan Passage).

II.2 Model Configuration and Boundary Conditions

The Model Configuration is the same as in the ReaL and use constant horizontal/vertical diffusivity coefficients which are the same as in the ReaL (tab. 1). The model is forced at the surface by an artificial heat flux balanced all along the simulation. This flux has a value of \(-800 \text{ W m}^{-2}\) over SA (and zero all over the rest of the domain) and it acts on this region in January. It is equal to \(\sim 1.61 \text{ W m}^{-2}\) all over the domain for the rest of the year in order to balance the heat loss over the SA. An estimation of the outflow of Deep Waters at Otranto Strait (not shown here) has shown a value of 0.2 Sv-0.3 Sv which fits the well known climatology [8]. Temperature and Salinity vertical profiles as well as zonal and meridional component of speed currents are prescribed at both boundaries: the model is forced for 120 months with the climatologies Medar-Medtlas II/MyOcean computed for the period 1987-1995 in order to reproduce (as stated before) the characteristics of EMT period. For the rest of the numerical experiments the model has been forced with the climatologies of the period 1996-2012 (Float data/MyOcean) in order to reproduce the most recent characteristics of Mediterranean Circulation. The same conditions have been considered for the wind fields. Two classes of experiments have been prepared: one with wind forcing (YWIND) and one without wind forcing (NWIND).

II.3 Effects of external forcing on the vorticity of the Ionian

Fig. 10 shows \(\zeta\) in the first 500m in the Ionian Sea between 16E-20E and 34.5N-38N for the YWIND and NWIND experiment. The mean value of \(\zeta\) is equal to \(-1.04 \times 10^{-7} \text{s}^{-1}\) and \(-0.68 \times 10^{-7} \text{s}^{-1}\) respectively. Despite the tendency towards more cyclonic conditions observed in the both time series, after the change in the Eastern Boundary conditions from Medar to Float data, no reversal has been observed. The same temporal length of this tendency (approximately 15 years) is larger with respect to the variability (4-10 years) of upper
layer circulation observed by [3][7]. Fig. 11 a shows the corresponding Hovmoeller diagram for the $\zeta$ in the Ionian Sea for the YWIND experiment: the bilayer is still present but the zero level for the vorticity moves upwards (up to 500m and less during the last 50 months of the experiments) after the changes in T-S conditions on the Eastern Boundary. This upward movement of zero level affects our computations of $\zeta$ introducing the tendency for this quantity to move towards less anticyclonic condition. This shows that eventual changes in T-S characteristics of LIW may affect the circulation of the Ionian. On the bottom the circulation is permanent cyclonic. Fig. 11 b shows the Hovmoeller diagram of the salinity in the SA during YWIND experiment: the clockwise circulation in the Ionian Sea lead to an enhanced import of fresher AW in the SA which produces a decrease of Salinity in the upper layers of the SA [7]. No changes in the salinity trends have been observed during the experiments confirming that no changes in the circulation took place in the Ionian. The absence of any reversal has been confirmed observing the maps (Fig. 12) of mean Sea Surface Height (hereafter SSH) overlapped with the streamlines of the velocity field in the first 100 m. For both period 1987-1995 (a,b) and 1996-2012 (c,d) the Ionian circulation is characterized by a permanent positive SSH signal which corresponds to a clockwise circulation. The circulation in the case of YWIND is stronger with respect to NWIND: this clearly derives from the action of wind over the basin. The input of vorticity estimated from wind stress curl is equal $-0.19 \times 10^{-5}$ s$^{-1}$. In both cases, the area of Gulf of Sirte is characterized by permanent positive values of SSH: this may allow to conjecture that the gyre present in the area is not directly wind-induced but, probably, is a topographic feature, eventually reinforced by wind.
Figure 12: Mean Sea Surface Height and mean velocity field in the first 100m for the period 1987-1995 and 1996-2012 for YWIND (a,c) and NWIND (b,d)
Fig. 13 a,b,c shows the behavior of a tracer between 3500 and 4000m released in the Cre- tan Passage during the period 1987-1995 of YWIND experiment. The tracer follows a cy- clonic path: entering the Ionian (a), moving along the eastern flank (b) and then filling the Ionian bottom layer (c). This reproduces the behavior of Deep Waters produced during the EMT in the Aegean Sea [7] [14]: these waters exiting via Kasos Strait propagated westward adjoining the Cretan Arc, mostly continuing up to the northern end of the Hellenic Trench near 37°N and finally spreading along the eastern flank of the Ionian following a cyclonic circulation.

Tab. 3 shows the vorticity budget in the case of NWIND experiment: the presence of two lateral open boundaries changes dramatically the budget of vorticity. Baroclinicity is not negligible anymore probably due to the advection in the system of water with different characteristics (AW and LIW) which gives rise to important gradient of density/pressure in the system. The main terms are the stretching and baroclinicity term. The combination of stretching and advection guarantees the negative sign in the vorticity field.

<table>
<thead>
<tr>
<th>Vorticity Budget Component</th>
<th>NWIND</th>
</tr>
</thead>
<tbody>
<tr>
<td>Advection</td>
<td>$-5.60 \times 10^{-5}$</td>
</tr>
<tr>
<td>Diffusion</td>
<td>$3.12 \times 10^{-5}$</td>
</tr>
<tr>
<td>Stretching</td>
<td>$-1.03 \times 10^{-4}$</td>
</tr>
<tr>
<td>Baroclinicity</td>
<td>$1.23 \times 10^{-4}$</td>
</tr>
</tbody>
</table>

Figure 13: Spatial distribution between 3500 and 4000m of a passive tracer in three different steps of the simulation during the YWIND experiment for the period 1987-2005.
III. Effects of the variability of Eastern Boundary Conditions on the vorticity balance of the Ionian Sea

In this section we discuss the effects of variability in the eastern Boundary conditions characteristics on the vorticity balance of the Ionian Sea.

III.1 Domain of integration

The domain of integration is the same as in the experiments in II, Fig. 9

III.2 Model Configuration and Boundary Conditions

The Model Configuration and boundary conditions are the same as in the experiments of section II in the first 120 months. For the rest of the numerical experiments the vertical profiles of Temperature and Salinity derived from Argo Float data for the period 1996-2012 prescribed at the Eastern Boundary have been corrected. 1 °C / 0.5 °C respectively have been added to the vertical profiles of Temperature. 0.25/0.15 psu have been subtracted from the vertical profiles of Salinity. These two class of experiments have been identified as Strong/Mild TS respectively. Wind and MyOcean forcings have not been modified. The reason for doing this correction comes from the necessity to restablish the SA as main source of deep waters for the Ionian. During NWIND and YWIND for the period 1996-2012 we employed tracers released simultaneously in the SA, Kythira strait and Cretan Sea. At the end of experiments we didn’t observe on the bottom of the Ionian any sign of the tracer (so of ADW) released in the SA while we observed the tracers released in the Kythira and Cretan Passage. This meant that deep waters formed in SA could not reach the bottom of Ionian and the deep layers of the basin were still under the influence of EMT. On contrary observations [4] [28] have shown that the ADW are again able to reach the deep layers of the Ionian. To restablish the primary role of the SA we decided to make lighter the vertical profiles of temperature and salinity prescribed at eastern Boundary.

Six experiments have been considered and will be indicated with exp1-exp6. The main characteristics (in terms of forcing) of these experiments are:


Finally an additional experiment (exp6) has been considered which is longer with respect to exp1-exp5. Exp6 is divided in three phases:

- first phase : 120 months forced with MyOcean, Wind Forcing and Medar/Medatlas TS conditions for the period 1987-1995
- second phase : 280 months forced with MyOcean, Wind Forcing and Strong TS for the period 1996-2012
- third phase : 100 months forced with MyOcean, Wind Forcing and Medar/Medatlas for the period 1987-1995
The passage in the TS conditions among the three phases takes place as in the exp5. Fig.14 shows the $\zeta$ in the exp1-5 in the first 500m of the Ionian Sea between 16E-20E and 34.5N-38N. In all the cases in correspondence of the change in TS conditions on the Eastern Boundary a complete reversal from anticyclonic to cyclonic circulation takes place and the system doesn’t go back to the previous circulation conditions. This change in sign of the vorticity is clear in fig.15 where the maps of mean SSH have been overlapped with the streamlines of the velocity in the first 100 m. The mean SSH changes sign between the two periods: a positive sign is present in the period 1987-1995 which corresponds to the clockwise circulation while a negative sign corresponding to a counterclockwise circulation is present in the period 1996-2012. Fig.16 shows $\zeta$ in the exp6 in the first 500m of the Ionian Sea between 16E-20E and 34.5N-38N. In this case two reversal take place: the system switches from clockwise to counterclockwise circulation with the establishment of STRONG TS condition and switches back to clockwise circulation with the re-establishment of Medar-Medatlas Conditions. The change in sign of vorticity in the Ionian is confirmed by the salinity behavior in the SA (fig.17). The hovmoeller diagram shows that the clockwise conditions (first phase) in the Ionian leads to an enhanced advection of AW in the SA which results in a decrease in the salinity of the basin. Then a counterclockwise circulation, in the second phase, enhances the advection of saltier LIW in the SA leading to an increase of salinity. Finally in the third phase a re-established clockwise conditions increase the advection of AW resulting in a new decrease of salinity in the SA. Still the hovmoeller diagram of $\zeta$ shows the temporal evolution of $\zeta$ in the Ionian (fig.18). In the second phase a positive vorticity layer occupies completely the first 500m of the Ionian. The intermediate layer up to 3000 m are occupied by a negative vorticity layer. The bottom layers are still positive in sign (fig.18). This vertical structure in the second phase may be explained with the filling of the intermediate layers with less dense waters with respect to the Medar-Medatlas conditions which are able to invert the sign of the components in the vorticity budget.
Comparing tab. 4 with tab.3 (NWIND experiment) the strong differences between the two situations are clear: stretching and baroclinity are still the main terms but in the exp6 second phase stretching is the dominant term with a positive sign. The input of vorticity from the wind is the same as in the YWIND experiment second phase (\(-0.14 \times 10^{-5} \text{s}^{-1}\), not reported in tab. 4) and will be analyzed in the energy section. Despite this reversal could be attributed to this strong changes in the TS conditions on the Eastern Boundary, smallest variations in TS with respect to the Medar-Medatlus conditions can induce a reversal themselves. In fact fig.19 shows the temporal evolution of SSH and the velocity fields in eight months in presence of linear increase/decrease in T/S values from the Medar-Medatlas to Strong TS conditions: after 6 months (c) Ionian is mostly occupied by a negative SSH value which corresponds to a counterclockwise circulation. This corresponds to an estimated increase of \(~0.2^\circ\) C in the temperature and a decrease of \(~0.08\) psu for the salinity. This range of values have been confirmed recently with experiments with more complex forcing applied to AIS (Querin S., personal communication).
Figure 17: Hovmoeller diagram of the salinity (in psu) in the Adriatic during the exp6

Figure 18: Hovmoeller Diagram of $\zeta$ (in s$^{-1}$) in the Ionian Sea during the experiment exp6

Table 4: Vorticity budget in the second phase of exp6: cumulated input (in s$^{-1}$)

<table>
<thead>
<tr>
<th>Vorticity Budget Component</th>
<th>NWIND</th>
</tr>
</thead>
<tbody>
<tr>
<td>Advection</td>
<td>$-1.37 \times 10^{-4}$</td>
</tr>
<tr>
<td>Diffusion</td>
<td>$-1.10 \times 10^{-4}$</td>
</tr>
<tr>
<td>Stretching</td>
<td>$5.54 \times 10^{-4}$</td>
</tr>
<tr>
<td>Baroclinicity</td>
<td>$-2.90 \times 10^{-4}$</td>
</tr>
</tbody>
</table>
IV. Energetics of the AIS

In this section we discuss the effects of variability in the eastern Boundary conditions on the components of Energy Budget of the AIS system in the exp1-6.

IV.1 Energy Budget

Fig.20 and fig.21 show respectively the total Kinetics Energy (KE) and $APE_{QG}$ for the exp6. From both figures is clear that clockwise and counterclockwise circulation correspond to two different energetic states for the system. In fact the clockwise circulation corresponds to a lower energy state : approximately $4.26 \times 10^{14}$ J for the KE and $2.57 \times 10^{16}$ J for the $APE_{QG}$. These values correspond to some estimations (R.Mosetti, personal communication) based on QG box model for the AIS. The counterclockwise circulation corresponds for the system to a higher energy state : an approximately double value for the KE ($7.37 \times 10^{14}$ J) and one order of magnitude more for the $APE_{QG}$ ($2.64 \times 10^{17}$ J).
The main source for KE for the AIS are (eq.(12)) wind power $G_{KE}$ and $C(APE_{QG}, KE)$. Fig 22 shows the input of energy from wind over the AIS in the exp6. This figure shows the behavior of wind with respect to the variation of vorticity: while during the clockwise circulation wind reinforces the circulation [15] with an average input of $1.47 \times 10^8$ W, during the transition from anticyclonic to cyclonic circulation wind works against the circulation. So the wind is not responsible for the observed change and neither for the first peak in the KE.
observed in fig. 20. After the transition the average input of energy over the AIS lowers to $5.31 \times 10^7$ W. Concerning the $C(APE_{QG}, KE)$ (fig.23) it is the main source of energy for the AIS: the average input of $C(APE_{QG}, KE)$ during the period 1987-1995 is approximately $1.26 \times 10^9$ W while this value doubles in the period 1996-2012 ($2.78 \times 10^9$ W). The importance of this term in the KE equation appears clear in fig. 23 where we can observe the two peaks of $C(APE_{QG}, KE)$ which correspond to the two peaks of KE in fig.21. Still the second peak which corresponds to the transition from cyclonic to anticyclonic results more energetics not only because $C(APE_{QG}, KE)$ but because the sign of wind stress curl corresponds to that of transition adding more KE to the transition itself (fig. 22). The same considerations hold for the exp 1-5 (fig. 24, fig.25).

The main source term of $APE_{QG}$ is (eq.(13)) $G_{APE_{QG}}$. Fig 26 shows the buoyancy fluxes over AIS: the mean contribution is equal for all the experiments to $-5.79 \times 10^5$ W and is not able to explain the variability observed in $APE_{QG}$ and $C(APE_{QG}, KE)$ values. Our domain has two boundaries: AIS receives an input of $APE_{QG}$ from both Sicily Channel and Kythira Strait/Cretan Passage. In the first case for the period 1987-1995 we estimated an inflow of $APE_{QG}$ of about $1.67 \times 10^7$ W m$^{-3}$. The inflow from Kythira Strait/Cretan Passage is one order of magnitude greater: the average inflow estimated is $\sim 1.05 \times 10^8$ W m$^{-3}$. So the Eastern Boundary represents an important source of $APE_{QG}$ for the AIS. The inflow from Eastern Boundary (the conditions on the Western Boundary are maintained constant along the simulation) increase of 3 order of magnitudes during the TS stronger 1996-2012 conditions. This has been a consequence of changes in TS profiles: in fact the change in the inflow of $APE_{QG}$ has been still observed also in the case of constant zonal flow. Definitely the Levantine basin is an important source of $APE_{QG}$ for the AIS with respect to buoyancy fluxes and Sicily channel inflow.

Figure 24: Wind Power in Watt for the exp1 (red line), exp2 (blue line), exp3 (fuchsia line), exp4 (green line), exp5 (yellow)
Figure 25: Conversion term between KE and APE$_{QG}$ in Watt during exp1 (red line), exp2 (blue line), exp3 (green line), exp4 (fuchsia line), exp5 (yellow)

Figure 26: Buoyancy flux over the AIS in W during the exp1-exp5

IV. CONCLUSION

This study has been performed with the final purpose to provide a deeper insight into the dynamics of AIS. An approach based on an increasing complexity in the model forcing and domain has resulted in a more extensive understanding of AIS dynamics. In particular it has been shown that the influence of the Adriatic Sea outflow on the Ionian Sea dynamics is not affected by the bathymetry of the basin, resulting in both cases in a bi-layer structure for vorticity in the Ionian Sea induced by vortex stretching mechanism. On the contrary the vorticity and energy balance of our system appeared to be more influenced by the Levantine Basin, showing that AIS cannot act as a bipolar oscillating system (as suggested in the BiOS mechanism) but as a multipolar system. The reversal from anticyclonic to cyclonic circulation is strictly correlated with the inflow of APE$_{QG}$ from the eastern boundary and the rate of this change in the Ionian vorticity is correlated with rate of change of TS conditions in the area. This variation in the vorticity appears to affect the upper and intermediate layers of the Ionian and its northern-central part. The bottom lay-
ers are characterized by a permanent cyclonic circulation while the Southern Ionian (mainly the area of Gulf of Sirte) is characterized by a permanent clockwise circulation, eventually reinforced by wind. In this framework also the role of the wind appears to be marginal: it is able to reinforce/weaken the circulation but it is not able to change the sign of the circulation. The impacts of Temperature-Salinity profiles in the eastern basin over the Ionian circulation in 1995, 2006 and 2011 appear to be confirmed by the analysis of float data and the data from German cruises taken in these areas [28]. In fact significant changes in the potential density profiles (fig.27) of these water have been observed mainly at level of surface/intermediate layer (up to 500m with difference in the potential density ranging from 0.03 kg m$^{-3}$ to 0.3 kg m$^{-3}$) in those years apparently confirming that the source of variability of the Ionian circulation is represented by the Levantine Basin (as originally thought [3]) not by the SA thermohaline processes [7] [11] or by the wind [15] [23] [25].

**Figure 27:** Vertical profiles of potential density (in kg m$^{-3}$ in the Cretan Passage: a) 1995 (red line; anticyclonic state in the Ionian) - 1999 (blue line; cyclonic state in the Ionian); b) 2005 (blue line; cyclonic state in the Ionian) - 2006 (red line; anticyclonic state in the Ionian); c) 2010 (red line; anticyclonic state in the Ionian) - 2011 (blue line; cyclonic state in the Ionian).
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References


Chapter 4

Conclusions

The variability in the Ionian upper layer circulation which reverses on decadal scale from anticyclonic to cyclonic circulation and vice versa has been recently subject of a challenging debate in order to identify the mechanism(s) driving these reversals. In order to challenge all the contrasting hypotheses related to AIS dynamics and to rank the relevant mechanisms governing it a modeling approach based on an increasing complexity of the model and forcings has been followed. The starting point has been to consider an isolated AIS forced only with heat fluxes and an idealized/real bathymetry. The final point has been to consider an open AIS with full forcings, namely lateral flows of LIW and AW and wind stress. The BCs at the lateral boundaries and at surface have been carefully chosen in order to reproduce the main processes taking place in the system: the deep waters formation in the SA, the input of energy from the wind, the lateral fluxes of LIW and AW in the Ionian basin.

Our modeling approach has shown that the influence of the Adriatic Sea outflow on the Ionian Sea dynamics is not affected by the bathymetry of the basin, resulting in both cases in a bi-layer structure for vorticity in the Ionian Sea induced by vortex stretching mechanism: negative up to 2000 m, positive up to the bottom of the basin. The vorticity and energy budget of the AIS appears to be more influenced by the Levantine Basin instead of the Adriatic Sea (as suggested in the BiOS paradigm). This variation in the vorticity affects the upper and intermediate layers of the Ionian and its northern-central part while the Southern Ionian (mainly the area of Gulf of Sirte) is characterized by a permanent clockwise circulation, which appears to be topography induced and eventually reinforced by wind. The bottom layers are characterized by a permanent cyclonic circulation which does not reverse. The reversal from anticyclonic to cyclonic conditions and vice versa appears to be strictly correlated with the input of $APE_{QG}$ from the Levantine basin into the AIS. In fact the generation of $APE_{QG}$ by buoyancy fluxes is not able to explain alone the variability observed in $APE_{QG}$ in the AIS. Still the variation in the KE in the AIS
cannot be explained in terms of wind forcing (since it works against the circulation in the period of transition from anticyclonic to cyclonic) but it appears to be strictly correlated with the conversion of $APE_{QG}$ in KE. The rate of the reversal itself is strictly correlated with the magnitude of $APE_{QG}$ inflow which depends on the Temperature-Salinity of waters in the Eastern Basin. The impact on the vorticity budget of thermohaline properties of waters entering into AIS from its eastern boundary seems to be confirmed by the analysis of potential density profiles of waters in the Cretan Passage during the reversals observed in 1997, 2006 and 2011. These vertical profiles show a change in the potential density values mainly at surface/intermediate layers which correspond timely to two different states of circulation in the Ionian Sea. Definitely the final outcomes of this work do not support the idea of a closed dipole AIS as reported in the BiOS mechanism but on the contrary the idea of multipole structure of AIS including the Cretan Passage and Cretan Sea.

New experiments are planned to include as eastern boundary conditions Temperature/Salinity profiles recently available for Cretan Sea/Passage. Incoming analysis will try to clarify the role of the interannual variability of wind stress forcing in shaping the interaction processes of this multipole structure and the decadal variability of the Ionian upper layer circulation. Another important issue to resolve is the understanding of feedback mechanisms and time delays between the variability of thermohaline properties of Levantine and Ionian waters. Finally, based on most advanced GCM hindcasts for the Mediterranean Area, an historical analysis will be done in order to identify previous reversals in the Ionian and the characteristics of wind forcing/thermohaline fluxes over the area during those periods.
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