A Sensitivity Study of the General Circulation of the Western Mediterranean Sea. Part II: The Response to Atmospheric Forcing

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ABSTRACT

This paper investigates the influence of sea surface thermohaline fluxes and wind stress on the circulation of the Western Mediterranean Sea using a high-resolution 3D primitive equation model. An 18-year experiment was forced with the daily output of a fine grid mesh numerical weather prediction model. The major characteristics of the circulation are well reproduced. The basin surface circulation is cyclonic over all of the basin. The two anticyclonic Alboran gyres are present. The instabilities of the Algerian Current generate large anticyclonic eddies that invade the whole Algerian Basin. The Liguro-Provençal-Catalan Current is well marked. Deep water convection down to the bottom only occurs during the first 3 years, then winter intermediate water is produced. The north-south gradient of the atmospheric thermohaline fluxes induces a northward surface transport of water from the Algerian Basin into the Liguro-Provençal Basin. This pattern can be associated with the Balearic front. Sensitivity experiments show that the wind stress curl reinforces the cyclonic circulation of the Liguro-Provençal Basin through a Sverdrup balance mechanism and contributes to deep-water formation. It is also suggested that the variations of the transport in the Corsican Channel are linked to the wind stress action rather than the heat flux gradient between the Tyrrenian and Ligurian Seas.

1. Introduction

The Mediterranean Sea is a thermodynamic system that transforms the Atlantic incoming oceanic waters into denser ones through processes dependent on interaction with the atmosphere. The difference of density between the Western Mediterranean and the Atlantic Ocean and between Western and Eastern Mediterranean basins drives the mean transports through the Straits of Gibraltar and Sicily. They both contribute to the forcing of the cyclonic circulation of the modified Atlantic water (MAW hereinafter) and the Levantine Intermediate Water (LIW hereinafter) (Lacombe and Tchernia 1972; Millot 1987, 1991; Herbaut et al. 1996).

The salinity of Atlantic water entering through the Strait of Gibraltar is about 36.6 psu. Its temperature varies seasonally with a maximum of 21°C in August and a minimum of 17°C in March (Lacombe and Richez 1982). It is fresher and lighter than the outflowing Mediterranean Water, which has a salinity of 38.25 psu and a temperature of 13.30°C. Usually the Atlantic water inflow forms two anticyclonic gyres in the Alboran Sea (Lanoix 1974; Tintore et al. 1988). As recently observed on thermal infrared satellite imagery, these can show vacillations with a period of a few weeks (Heburn and La Violette 1990). As the MAW emerges from the Alboran Sea, it flows from the Spanish coast (from 2°W) toward the Algerian coast (1°W) and quite frequently gives rise to an intense frontal jet, known as the Almeria–Oran front (Tintoret et al. 1988). It then forms the Algerian Current, which flows eastward along the African coast. It generally becomes unstable around 1°–2°E forming cyclonic and anticyclonic eddies and meanders through baroclinic instability (Millot 1985; Mortier 1992; Beckers and Nihoul 1992). These eddies drift eastward along the African coast at velocities of a few centimeters per second, but only anticyclonic eddies grow up to a diameter of 100 km. They sometimes detach from the coast and then slowly migrate into the

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open sea, forming the quasi-steady eddies identified for periods of several months from infrared satellite data. They may span more than 200 km in diameter and eventually return to interact with the stream (Taupier-Letage and Millot 1988). Part of this MAW penetrates into the Eastern Mediterranean Sea compensating the water mass lost by evaporation (Bethoux 1979, 1980), the other flows along the northern coast of Sicily, the coast of Italy, and passing through the Strait of Corsica, contributes to the feeding of the Ligurian–Provençal–Catalan Current.

The northern part of the basin is characterized by the Ligurian–Provençal–Catalan Current (the northern current hereinafter), which flows cyclonically along the coasts of Italy, France, and Spain. Its transport is of the same order of magnitude as the incoming Atlantic water transport at Gibraltar (Lacombe and Tchernia 1972; Lacombe et al. 1981; Bethoux et al. 1982; Taupier-Letage and Millot 1988). It displays a seasonal variability with a maximum transport (1.6 Sv, Alberola et al. 1995) in winter and a mesoscale variability marked by the growth of meanders that breakdown into eddies on a smaller scale than these generated by the Algerian Current. The deep-water formation occurring in the MEDOC region (MEDOC Group 1970) reinforces this circulation (Crépon et al. 1989; Madec et al. 1991a,b). The northern current then follows the continental slope of the Balearic Sea where it has been studied by Font et al. (1988), Font (1990), and Pinot et al. (1994). A branch of the surface flow is deflected to the northeast by the Balearic Islands, while the main branch continues southward along the Spanish coast and reenters the Algerian Basin (Millot 1987) contributing to reinforce the surface cyclonic circulation.

In winter and in the northern basin, this MAW is transformed into winter intermediate water (WIW hereinafter—the so-called Riviera Water in Lacombe and Tchernia 1960) characterized by a minimum of temperature (Millot 1985), into Western Mediterranean Deep Water (WMDW) in the Western Mediterranean (MEDOC Group 1970; Stommel 1972; Gascard 1978), and into the Levantine Intermediate Water (LIW) in the Eastern Mediterranean Sea. The circulation of the interior Mediterranean water masses (WIW, LIW, WMDW) is not as well known as the path of the MAW.

Observations (Gascard and Richez 1985; Benchhora and Millot 1995a; Pinot et al. 1994) show the presence of WIW in the Balearic Sea, in the Algerian Basin, and the Alboran Sea, but these observations are still too sparse to define a circulation scheme. LIW is characterized by maximum temperature and salinity values ranging from $\theta = 14.10^\circ$C and $S = 38.75$ psu in the Sicilian Channel to $\theta = 13.25^\circ$C and $S = 38.50$ psu at the Strait of Gibraltar (Lacombe and Tchernia 1972). According to Millot (1987), the LIW coming from the Strait of Sicily into the Western Mediterranean turns to the right along the Italian coasts. Part of this water passes through the Corsican Channel (Astraldi and Gasparini 1992), but the major part continues to flow cyclonically around the Tyrrhenian Sea toward the Sardinia Channel. It then runs northward along the western coasts of Sardinia and Corsica (Perkins and Pistek 1990) and joins with the LIW emerging through the Corsican Channel. The LIW then flows along the French and Spanish coasts. Recent observations (Send et al. 1995) confirm this cyclonic path of LIW. Coastal-trapped waves and topography seem to be a major factor in driving the LIW flow. Large eddies occurring in the Algerian Basin might disrupt the LIW circulation (Millot 1987). As for the WMDW circulation, it has never been investigated and its path is still conjecturable.

The mechanisms that drive the surface circulation have been addressed by many authors. In Part I of this two part paper, Herbaut et al. (1996) showed that the forcing due to the horizontal density gradients at the straits of Gibraltar and Sicily are able to generate a cyclonic circulation of the Atlantic Water. Indeed, according to Tintore et al. (1988), Perkins et al. (1990), and Speich et al. (1996), the circulation in the Alboran Sea and along the Algerian coast is mainly driven by the density gradient between the Atlantic and the Mediterranean waters. Numerical process studies (Madec and Crépon 1991) showed that deep-water formation induced by thermohaline forcing reinforces the northern current. However, violent winds encountered in winter may also enhance the circulation. Heburn (1994) and Pinardi and Navarra (1993) argued that the winds contribute to drive the general cyclonic circulation. According to Madec et al. (1996), winds favor deep-water formation preconditioning in the Gulf of Lion. Besides, Krivosheya (1983) suggested that the winds are responsible for the circulation over the Tyrrhenian Sea.

The present study aims at investigating the role of the different forcings on the circulation of the western basin by examining a series of numerical studies done with a high-resolution ocean general circulation numerical model. OGCMs are now accurate enough to provide a coherent and quantitative description of the circulation allowing us to understand the mechanisms driving the exchange between the subbasins and the paths of the water masses. Previous numerical studies devoted to the whole Mediterranean Sea (Zavatarelli and Mellor 1995; Roussenov et al. 1995; Pinardi and Navarra 1993; Wu and Haines 1996; Haines and Wu 1995) reproduce the main features of the basin-scale circulation of the Western Mediterranean. However, because of their coarse horizontal resolution, they do not adequately resolve the coastal currents, which are of fundamental importance in the dynamics of the western basin (Herbaut et al. 1996).

The paper is organized as follows: after a brief presentation of the numerical model (section 2), we discuss the results of the basic experiment E1, which included all the forcings (section 3). Attention is focused on the circulation variability and the ensuing water mass distribution. We then perform sensitivity studies to inves-
Table 1. Synthesis of the different numerical experiments.

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FIG. 1. Initial potential density profiles for the “idealized” Atlantic Ocean (dashed–dotted line), the Eastern Mediterranean (dashed line) Basin, and the Western Mediterranean Basin (plain line), in $\sigma_t$ units.

Table 1. Synthesis of the different numerical experiments.

- **Forcing/experiment:**
  - E1
  - E2
  - E’2
  - E3

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3. The basic experiment (E1)

In this section, we address the combined effect of straits, thermohaline, and wind forcings on the circulation and the formation and paths of the water masses in the Western Mediterranean Sea.

The model was forced with daily atmospheric thermohaline fluxes, wind, and density gradients at the Straits of Gibraltar and Sicily. The simulation was run for an 18-year period to simulate the LIW advection from the Strait of Sicily to the Strait of Gibraltar. Since attention is focused on the ocean dynamical processes driving the circulation, we used a perpetual year forcing to facilitate the analysis of the ocean response by avoiding its contamination by atmospheric interannual variability.

Following Herbaut et al. (1996), the Western Mediterranean Sea was connected to two idealized basins representing the Gulf of Cadiz and the Ionian Sea. Each basin was initialized by horizontally constant vertical profiles of temperature and salinity representative of the studied area and obtained from the Levitus climatology (Levitus 1982) (Fig. 1): The only initial horizontal density gradients were located at the Straits of Gibraltar and Sicily. At $t = 0$, the two “dams” at the straits were opened. To maintain constant the forcing due to the density gradients between the Mediterranean Sea and the two adjacent basins, we restored the salinity and temperature profiles in the Gulf of Cadiz and the Ionian Sea toward temperature and salinity profiles from the Levitus climatology (monthly means for temperature and seasonal means for salinity).

a. The atmospheric forcing

Since the basin-scale circulation of the Mediterranean is dependent on mesoscale oceanic phenomena such as deep-water formation, which is linked to the pattern of
the mistral gusts (Madec et al. 1996), the atmospheric forcing must have adequate spatial and time resolutions, namely, less than 50 km and at least on a daily basis. The sole available atmospheric forcing presenting the above characteristics is provided by the PERIDOT model. PERIDOT is a numerical weather prediction model run in an operational mode by Météo-France over Europe and the Mediterranean area (Imbard and Joly 1986). The daily PERIDOT output are provided on a fine grid mesh (33 km × 33 km). Owing to the availability of the data when we started this study, we used the period ranging from August 1988 to July 1989, which can be considered as representative of a climatological year (i.e., there is no abnormal season).

The annual heat budget computed from this dataset corresponds to a sea surface heat flux of 0.15 W m⁻², which is lower (in term of sea surface heat loss) than that computed by Béthoux (1979) from the heat transport through the Gibraltar and Sicily straits, which was estimated to 7 W m⁻². The PERIDOT heat budget is unable to compensate the heat transport through the two straits. In order to avoid a long-term heating of the Western Mediterranean Sea and to stabilize the heat budget, a constant bias uniformly distributed in space was added to the atmospheric fluxes. To prevent local drifts in temperature of the model, the time evolution equation of the oceanic sea surface temperature includes a restoring term toward the PERIDOT SST. Freshwater flux is computed from latent heat flux. Precipitation and river runoff are parameterized in the salinity equation by imposing a restoring term to the Levitus climatology.

The temperature and salt restoring terms were applied during summer months only. In winter, this procedure would attenuate deep-water formation in the MEDOC region since the PERIDOT SST is too warm and the Levitus salinity too low compared to the observations (MEDOC Group 1970).

After 12 years, the salt content is stabilized and the heat content drift is lowered to an increase of 0.1% over 5 years (Fig. 2). The salt flux corresponds to an annual water loss of about 0.6 m yr⁻¹ and the heat loss is about 5.7 W m⁻² during the last year of integration.

We now focus our attention on the path of MAW, WIW, and LIW.

b. Characteristics and paths of the water masses

1) The volume transport in the Strait of Gibraltar

The annual mean transport in the Strait of Gibraltar stabilizes at 0.83 Sv, with a seasonal oscillation of 10%
FIG. 4. Experiment E1: $\theta$–$S$ diagrams in different areas of the Mediterranean Sea: The points indicate measurements from BNDO in the different areas, while the plain line is the mean diagram from the model’s simulation.
(Fig. 3). The transport is minimal from June to October with a value of about 0.78 Sv. The mean value is close to recent direct and indirect estimations, which range between 0.7 to 0.96 Sv (Bryden et al. 1989, 1994; Bryden and Kinder 1991; Harzallah et al. 1993; Bray et al. 1995). This value is of the same order as found in the numerical model of Wu and Haines (1996), higher than that of Roussenov et al. (1995), who showed a mean transport of 0.66 Sv, but far less than that of Zavatarelli and Mellor (1995), who found 1.35 Sv.

2) The path of the surface water

(i) Hydrology of the surface layer

In the Strait of Gibraltar, the simulated temperature and salinity of the inflowing Atlantic surface water (averaged over the top 100 m) vary respectively from 15°C (winter) to 17°C (summer) and from 36.75 psu (winter) to 36.65 psu (summer), the latter values being slightly higher than these observed by Gascard and Richez (1985) who measured a salinity between 36.2 and 36.5 psu in October. Comparisons between observed SISMER data (MAST/MEDATLAS Consortium 1996) and simulated θ–S diagrams show that the hydrology of the upper layer is correctly reproduced in the Alboran Sea and in the western Algerian Basin (Fig. 4). In the Tyrrhenian Sea, the simulated summer hydrology is in agreement with observations (MAST/MEDATLAS Consortium 1996) (Fig. 4), but the simulated winter temperature is too high. In the Liguro–Provencal Basin, the summer salinity is higher than the observed salinity and the situation reverses in winter. A north–south surface density gradient sets up after 18 years of integration: Modified Atlantic water (MAW) occupies the Algerian Basin while denser water is present in the northern basin (Fig. 5).

(ii) The basin-scale circulation

The simulated basin-scale surface circulation is cyclonic (Fig. 6). The Atlantic water is advected into the Algerian Basin and flows along the Algerian coast as an unstable coastal current generating large anticyclonic eddies. Part of the modified Atlantic water flow continues eastward and further divides, with one branch penetrating into the Tyrrhenian Sea and the other into the Strait of Sicily. In the northern basin, the northern current is identified from the Corsican Channel to Cap Creus at the Spanish–French border. This current is fed by the Eastern and Western Corsican Currents.

Significant water exchanges between the Liguro–Provencal and the Algerian Basins are found in the simulation. MAW flows northward along the west coast of Sardinia more or less regularly while Mediterranean surface water (defined by waters having a salinity greater than 38 psu) is advected by small mesoscale eddies from the Liguro–Provencal Basin into the Algerian Basin.

The southward transport of the Mediterranean surface water is weaker than the MAW northward transport. Its annual mean value is about 0.4 Sv. The northward flow of MAW is driven by the atmospheric thermohaline forcing, as shown in section 4. It is a significant source of freshwater (its salinity is less than 38 psu) for the Liguro–Provencal Basin and can contribute to inhibit the deep-water formation. Such a MAW flow from the Algerian Basin into the northern basin between Minorca and Sardinia has been suggested by Millot (1987) from observations.

The circulation in the different subbasins in the last year of the simulation is analyzed hereafter.

(iii) The Alboran Sea

The circulation vacillates between two states characterized by an eastward current forming meanders and anticyclonic eddies. In a first state, an anticyclonic gyre sets up in the western part of the basin (Fig. 7). When this gyre is fully developed, its diameter can span 120 to 150 km. The Atlantic water flows around it and forms two anticyclonic meanders along the African coast, each about 80 km in diameter. The first state appears mainly in June and the second at the end of January–beginning of February. The two states are likely to be shifted with a phase lag of 180°. The vacillation of the gyre does not seem linked to the transport variations in the Strait of Gibraltar.

The dimensions of the simulated anticyclone (corresponding to the first state, 100 km in diameter and
200 m deep) are in good agreement with in situ measurements (Gascard and Richez 1985; Tintore et al. 1991). In situ observations also show a temporary disappearance of the anticyclonic gyre in the western Alboran Sea and its replacement by a coastal current along Morocco (Cheney and Doblar 1982; Heburn and La Violette 1990). However, the observations show that the actual gyre disappears within a few days, while it takes a longer time in the simulation.

(iv) The Algerian Basin

The model circulation in the Algerian Basin is characterized by the presence of several anticyclonic gyres that evolve very slowly.

In the western part of the basin, we notice a quasi-stationary anticyclonic gyre centered around 1°E. This gyre, with a diameter of about 250–300 km, occupies the whole width of the basin between the Algerian coast
and the Balearic Islands (Fig. 6). The associated surface velocity is about 40 to 50 cm s$^{-1}$. MAW flows around this gyre before entering the central part of the Algerian Basin at the longitude of Algiers. According to this circulation pattern, the front between MAW and Mediterranean surface water is located south of Ibiza. This model circulation pattern does not correspond to observations, which usually show the front between Almeria and Oran and MAW flowing along the coast between Oran and Algiers (Millot 1985; Benzohra and Millot 1995a,b).

In the center of the Algerian Basin, the simulation shows the presence of some anticyclonic gyres with diameters that can reach up to 300 km. They may interact with the coastal current and deflect it offshore. They sometimes split into smaller eddies. Similar scenarios have been described by Taupier Letage and Millot (1988) and Benzohra and Millot (1995a,b) from satellite infrared imagery and in situ data. However, the observed gyres are smaller (with a diameter about 200 km) and seem to have an eastward drift, while the simulated gyres are quasi-stationary.

Consequences of the simulated circulation pattern in the Algerian Basin, that is, the absence of a well-defined coastal current and the presence of large anticyclonic gyres, are that the spreading of MAW may be higher than observed and that exchanges between the Algerian Basin and the Liguro-Provençal Basin are difficult to assess.

(v) The Strait of Sicily

The MAW progresses eastward and then separates into two branches at the entrance of the Strait of Sicily, two-thirds of which flow through the strait, one-third inflowing in the Tyrrhenian. Based on water and salt budgets, Béthoux (1980) estimated this ratio as $1/3$. The mean annual transport in the Strait of Sicily is about 0.65 Sv and presents significant seasonal variations (Fig. 3). It is close to the indirect estimations of 0.675 Sv from Harzallah et al. (1993) and the simulated values found by Haines and Wu (1995) and Wu and Haines (1996), whose mean value is 0.7 Sv, and Zavatarelli and Mellor (1995), which is 0.6 Sv. The direct estimations for the MAW transport are still controversial; they range from 0.3 Sv in August (Gelsi and Mosetti 1984), 1.5 Sv in January–February (Garzoli and Maillard 1979) up to 3.34 Sv in March (Manzella et al. 1988).

(vi) The Tyrrenian Sea

In the south of the Tyrrenian Sea, the simulated circulation is characterized by the presence of several variable mesoscale eddies. North of the Strait of Bonifacio, a cyclonic gyre (with a diameter around 80 km)
Fig. 9. Experiment E1: Horizontal velocity field at 37 m, filtered over the first 10 days of December in the 18th calendar year of integration.

is always present except in winter. Such a gyre has been observed throughout the entire year (Artale et al. 1994) and can be related to the wind stress curl (Perilli et al. 1995). The variations of the simulated transport in the Corsican Channel are in good agreement with the in situ measurements (Astraldi and Gasparini 1992) with a maximum in winter and a minimum in summer (Fig. 8).

(vii) The Liguro-Provençal Basin and the Balearic Sea

The simulated Western and Eastern Corsican Currents merge north of Corsica in the Ligurian Sea to form the northern current as observed by Béthoux et al. (1982), Millot (1987), and Astraldi et al. (1990). Its transport off Toulon is around 1.6 Sv, which is close to the above observations. In December, off Cap Creus, the northern current is deviated southward by the wind-driven cyclonic circulation that develops over the northern basin (see section 5). Therefore, the Catalan Current is no longer fed by the Liguro-Provençal Current and disappears (Fig. 9). In July, the Catalan Current appears again with a surface velocity of about 15 cm s$^{-1}$ (Fig. 6). However, Font et al. (1988) and Millot (1991) observed the continuity between the northern current and Catalan Currents year-round.

From February until July, the simulation shows a 0.35 Sv transport of MAW from the Algerian Basin into the Balearic Sea through the Ibiza Channel. This MAW is not advected in the interior of the Balearic Sea, but rather flows along the northern Balearic coasts with a thickness of about 60–80 ms. The transport is, however, insufficient to reproduce the North Balearic front as observed by Pinot (1994). During the rest of the year, the surface flow is directed southward and Mediterranean surface water enters the Algerian Basin through the Ibiza Channel. According to observations from Millot (1987), anticyclonic eddies would indeed be able to advect MAW from the Algerian Basin into the Balearic Sea. Hydrological sections performed in May 1990 (Pinot et al. 1994) show a 1 Sv MAW inflow through the Ibiza Channel.

3) Deep-water formation

Particular attention was paid on the transformation of surface water into deep water. Convection associated with deep-water formation is modeled by a TKE scheme providing a high vertical diffusion coefficient when the stability is small. During the first three years, deep-water formation occurs in an area centered at 41°30′N, 5°E whose diameter is about 100 km (Fig. 10). The convection reaches a depth of 1500 m at the end of February. The temperature of the newly formed deep water is 12.80°C and its salinity 38.42 psu (Fig. 10). Deep water stops forming in the following years. The surface water is not salty enough (the surface salinity maximum is less than 38.35 psu) to reach a density of 29.10 σ$\text{s}$ units necessary to convect to the bottom. Two reasons can explain this salt content deficiency. First, the volume of LIW (characterized by a salinity higher than 38.5 psu in the northern basin) decreases: LIW was consumed during deep convection events during the first 3 years and has not yet been replaced by “new” LIW very slowly advected from the Strait of Sicily. Second, fresh-
FIG. 10. Experiment E1: Meridional temperature (in °C, upper), salinity (in psu, middle), and density (in σt units, lower) vertical sections at 4.6°E, time-averaged over the first 5 days of February in the 3rd calendar year of integration, from 0 to 1500 m.

FIG. 11. Experiment E1: Zonal temperature (in °C) vertical section at 42°N from the Spanish coast (3°E) to 5°E, time-averaged over the first 10 days of March in the 18th calendar year of integration. Contour interval is 0.1°C. Winter intermediate water is formed over the shelf in the Gulf of Lion. WIW that was formed farther north and has sunk down to 400 m is advected along the shelf as seen on the local temperature maximum east of 4°E.

er water (MAW) is advected from the Algerian Basin in the surface layer and the local evaporation does not compensate for the associated decrease in surface salinity.

4) WINTER INTERMEDIATE WATER

In the simulation, winter intermediate water (WIW) formation occurs in different places. Water colder and less salty ($T < 10°$, $S < 38.15$ psu) than deep water forms over the shelf in the Gulf of Lion. This water sinks along the continental slope to stabilize between 300 and 400 m and then flows southwestwardly along the Spanish coast (Fig. 11). In situ measurements have described this phenomenon, the water formed over the shelf stabilizing around 350 m (Fieux 1972). In the center of the Balearic Sea, at the beginning of March, the stratification in the simulation vanishes between 0 and 300 m and intermediate water ($T = 12.90°$ and $S = 38.20$ psu) forms. There are no in situ observations at this location to validate the simulation. In the Liguro–Provençal Basin between 40° and 42°N, 5°E and 6°E winter convection reaches 450 m in the simulation, with a water temperature of $12.80°$ and a salinity around $38.30$ psu (Fig. 12). Over the Iberian Shelf, dense water ($σ_t = 29.10$, $T = 12°$, $S = 38.30$ psu) forms at the end of the winter (Fig. 13). This formation may be due to the lack of freshwater runoff from the Ebro River since no such dense water has ever been observed in this region.

From March to June, the simulated WIW flow enters the Algerian Basin through the Ibiza Channel. At 230 ms, its velocity is about 7 to 8 cm s$^{-1}$ and the southward
transport between 200 m and the bottom reaches a maximum of 0.9 Sv in May (Fig. 14). Mesoscale activity can significantly modify this path; in June, an anticyclonic eddy, which sets up north of the Ibiza Channel, deflects the WIW current toward the Mallorca Channel. The southerly transport of WIW from the northern basin toward the Algerian Basin stops in September.

This WIW distribution is in agreement with in situ observations. Guibout’s atlas (1986) shows cold water ($T < 13^\circ$C) below 200 m in the Ibiza Channel. Benzhora and Millot (1995a) observed WIW (characterized by temperatures between $12.7^\circ$ and $13.1^\circ$C) in the western part of the Algerian Basin.

5) Path of Levantine Intermediate Water

(i) The Tyrrenhian Sea

In the model, LIW initially lies between 400 and 500 m, its temperature and salinity being $13.4^\circ$C and 38.52 psu. After the 18-year period of integration, the 300–600-m layer of the Tyrrenhian is filled with LIW inflowing from the Strait of Sicily, with temperature and salinity higher than $14^\circ$C and 38.70 psu (Fig. 15).

In situ observations (Guibout 1986) show water with a temperature lower than $13.6^\circ$C. This observed low temperature may result from vertical mixing of LIW (with temperature higher than $14^\circ$C) with underlying deep water (Hopkins 1988) or from horizontal mixing LIW with WIW coming from the Algerian Basin through the Sardinian Channel (Manzella et al. 1988). In situ observations in the Sardinian Channel show the presence of a water mass between 200 and 500 m whose temperature is lower than $13.3^\circ$C (Guibout 1986; Manzella et al. 1988), indicating that the second hypothesis is more appropriate. In the simulation, no eastward cold water flow is found between 200 and 700 m in the Sardinian Channel and therefore LIW can fill the Tyrrenhian Sea intermediate layer. The hypothesis that vertical mixing can lower the temperature at intermediate depths is not validated in our simulation.

(ii) The Algerian Basin

In the simulation, a LIW stream 100 km wide flows along the western coast of Sardinia over the shelf between 300 and 900 m. The salinity maximum is higher than 38.70 psu (Fig. 16). This water mass is slowly advected northward (about 1 cm s$^{-1}$). The existence of such a LIW stream is in agreement with in situ observations (Perkins and Pistek 1990). Part of the LIW flowing through the Sardinian Channel is also advected by mesoscale eddies in the center of the Algerian Basin, as proposed by Millot (1985) from in situ observations.

(iii) The northern basin

In the simulation, LIW enters the northern basin along the western coasts of Sardinia and Corsica, as observed by Miller et al. (1970) and Astraldi et al. (1990). There is no LIW flow through the Corsican Channel as described by Bethoux et al. (1982) and Astraldi and Gasparini (1992). Along the western coast of Corsica, the modeled LIW stream is located between 350 and 700 m with a salinity maximum of 38.64 psu at 500 m. As LIW is advected westward along the French coast, the thickness of the vein and its salinity maximum decrease. In the center of the northern basin, the upper limit of LIW is at 420 m and the salinity maximum is less than 38.52 psu. In the Ligurian Sea, in situ data show the existence of LIW around 200–300 m along the coast (Alberola 1994), which is shallower than in the model. In the Balearic Sea, LIW is only noticeable in the simulation around 600 m and the salinity maximum has further decreased to 38.46 psu. The observed LIW sa-
linity is higher and at shallower depth, with 38.50 psu at 300 m (Pinot 1994).

Sensitivity experiments were run to investigate the major mechanisms responsible for the driving of the circulation of the western basin of the Mediterranean Sea. Attention is now focused to the understanding of the relative contribution of the thermohaline and the wind forcings.

4. Thermohaline atmospheric forcing (experiments E2, E'2, E'2')

A sensitivity experiment E2 (with the same initial conditions as in experiment E1) was forced with daily sea surface thermohaline fluxes and horizontal density gradients at the Straits of Gibraltar and Sicily. Deep water convection does not occur since the initial density field is horizontally homogeneous and does not allow a correct representation of the mesoscale cyclonic circulation characteristic of the preconditioning phase in the MEDOC area.

We then ran a second experiment (E'2) where we parameterized the deep-water formation preconditioning phase by artificially increasing the PERIDOT atmospheric thermohaline fluxes in the MEDOC area by a factor of 4 during January and February. Convection then occurs down to the bottom at the end of February as in E1. This demonstrates that the wind plays an important role in the preconditioning phase of deep-water formation since in the experiment E1 where the wind is present, it was not necessary to artificially increase the thermohaline fluxes for generating deep convection. The circulation in E'2 is cyclonic at the surface and...
Fig. 15. Experiment E1: Zonal temperature (in °C, upper) and salinity (in psu, lower) vertical sections at 40.5°N across the Tyr-rhenian Sea, time-averaged over the first 10 days of March in the 18th calendar year of integration. Contour intervals are, 0.1°C and 0.05 psu.

reverses at depth, with velocities of about 2 cm s⁻¹ at 1000 m (Fig. 17) and somewhat contributes in reinforcing the surface cyclonic circulation in the Liguro–Provençal Basin in agreement with the analysis of Crépon et al. (1989).

A significant surface transport of MAW is found in E′2 between the Algerian and the Liguro–Provençal Basins, reaching 0.8 Sv in the upper 100 m between Minorca and Sardinia. It can be associated to the Balearic front (Deschamps et al. 1984) and to the northward current flowing along the west coast of Corsica as observed during the DYOME experiment (Taupier-Letage and Millot 1988). This transport is forced by the north–south density gradient through a mechanism as described by Wajisowicz and Gill (1986). In winter, the averaged density of the upper layer (0–100 m) in the southern part of the Algerian Basin is 27.8 σ₀ units, while it reaches 28.8 σ₀ units in the central part of the northern basin. The north–south density gradient is due to the heat fluxes gradient between the two subbasins, which can reach 100 W m⁻² in winter, and to the inflow of light Atlantic water through the Strait of Gibraltar. The northward MAW flow between Minorca and Sardinia may contribute to the inhibition of deep-water convection as in experiment E1 by advecting fresh and consequently light waters into the convection zone.

In November and December, a local maximum heat loss is observed in the PERIDOT data in the western part of the Gulf of Lion, in the vicinity of Cap Creus. Results from E′2 show an important cooling with an ocean surface temperature decreasing from 17°C at the beginning of November down to 13.2°C at the end of December. The surface density locally increases with gradients parallel to the coast. Vertical mixing occurs in the upper 250 m in the coastal water and a cold water tongue forms perpendicular to the coast. At the same time, the southward surface transport in the Ibiza channel decreases, while the transport between 200 m and the bottom increases (Fig. 18). A specific experiment (E′2) was thus designed to understand the mechanism responsible for this transport modulation. The model was forced with the winter (October to February) daily thermohaline atmospheric fluxes. The Straits of Gibraltar and Sicily were closed to suppress the basin-scale cyclonic circulation linked to the strait forcings. At the end of February, convection occurs down to 300 m in the Gulf of Lion as in E′2. A northward surface coastal current is established in the Balearic Sea while a southward counter-current appears at 450 m (Fig. 19). This current is associated with a baroclinic coastal jet generated by the thermohaline forcing as shown in Madec and Crépon (1991). The baroclinic character of the coastal current explains the opposite variations of the surface and deep transports in the Ibiza Channel as found in experiment
E'2. The convection area can thus be considered as a "source" for the deep current and a "sink" for the surface current. This can explain the propagation of the temperature signal observed in the Balearic Sea along the coast of Spain by Send et al. (1996).

5. Wind stress forcing (experiment E3)

Ovchinnikov (1966) suggested that the surface cyclonic circulation and the seasonal variability in the Western Mediterranean Sea is mostly driven by the wind. Krivosheya (1983) confirmed this hypothesis. In a low-resolution primitive equation model (¼° for the horizontal and 8 vertical levels), Pinardi and Navarra (1993) demonstrated the impact of a monthly wind in the formation of the cyclonic circulation in the northern basin. Heburn (1994) using a 3-layer model claimed that the wind is the major forcing for the cyclonic circulation. According to Madec et al. (1996), the wind plays an important role during the preconditioning phase of deep-water formation by generating a local cyclonic circulation. Comparison
between experiments E1, E2, and E’2 leads to the same conclusion.

A specific numerical experiment where the model was forced only by the wind was thus designed to study the impact of the wind on the circulation. The experiment E3 was forced during a 3 1/2-yr period with daily winds (on a perpetual year basis) by the annual PERIDOT wind stress during a 2-yr period in order to reach an equilibrium state, which was diagnosed by the time evolution of the kinetic energy. North of a line extending from the French–Spanish border (Cap Creus) to the south of Sardinia, the annual mean wind stress curl is positive (Fig. 20), with a northward gradient and a maximum in the Ligurian Sea. A strong negative wind stress is present in the western part of the Gulf of Lion.

A cyclonic circulation sets up in the Liguro–Provencal Basin and forms a well-defined coastal current along the French and Italian coasts contributing to reinforcing the Liguro–Provencal Current (Fig. 21). This current is fed by the merging of the Eastern and Western Corsican Currents. Its surface velocity is about 35 cm s\(^{-1}\) and the associated barotropic transport is 1.5 Sv. The cyclonic cell observed in the Ligurian Basin is due to a Sverdrup balance associated with the positive wind stress curl observed in this region. The part of the Liguro–Provencal Current forced by the wind can be interpreted as a planetary boundary layer current of Stommel’s type (1957). In Stommel’s model, a planetary boundary current is obtained only on the western coast of the ocean because the wind stress curl equals zero at the northern and southern boundaries. If the wind stress curl were nonzero at these two zonal boundaries, a coastal current wider than the western boundary current would be generated along them. The wind stress curl at the Italian and French coasts of the Liguro–Provencal Basin is relatively strong and induces a boundary current. The analysis of the different terms of the barotropic stream-function equation shows that the dissipation terms dominate in the coastal current, which confirms its “coastal boundary current” nature. The angle of the coast tends to reinforce the current and to narrow it as shown in Neumann and Pierson (1966, p. 223). A process experiment with a simplified geometry illustrates that these simple characteristics of the wind stress (wind stress curl not equal to zero at the northern coast) and of the geometry of the basin (the angle of the northern coast) are able to explain the formation of the Liguro–Provencal Current (Fig. 22). Thus, it appears that the wind contributes to increasing the Liguro–Provencal Current, already forced by the density gradient at the Straits of Gibraltar and Sicily (Herbaut et al. 1996).

Off Toulon, the cyclonic gyre is reinforced in winter and at the beginning of spring, which may play a significant role in the preconditioning phase of deep-water formation (Fig. 23). Furthermore, its transport fluctuations mirror the ones in the Corsican Channel. The variations in transport through the Corsican Channel are very similar to that estimated from current meter data collected during the same period (Astraldi and Gasparini 1992), the transport computed in the model being slightly higher (Fig. 23). According to this experiment, it is suspected that the transport variations in the Corsican Channel are linked to the variations of the wind stress rather than the heat flux gradient between the Ligurian and the Tyrrenhenian Seas, as suggested by Astraldi and Gasparini (1992). If this transport were driven by density gradients linked to heat fluxes, the current in the Corsica Channel would be baroclinic as in the Strait of Gibraltar, as was found by Herbaut et al. (1996) and Speich et al. (1996). Measurements of Astraldi and Gasparini (1992) show that this transport is mainly barotropic.

In the Balearic Sea and in the northwestern part of the Algerian Basin, the simulated circulation shows an anticyclonic circulation centered on the Balearic Islands. It flows in opposite direction to the observed circulation (Font et al. 1988) and to the flow generated by density gradients at the Straits of Gibraltar and Sicily. Since the wind generates a northward surface transport in the Ibiza and Mallorca Channels, it may be considered as responsible for the inflow of Atlantic water into the Balearic Sea as observed at some periods (Pinot et al. 1994). In the rest of the basin, the circulation is less intense. In the Tyrrenhenian Sea, the presence of a cyclonic gyre in the north and of a second cyclonic cell in the south is in qualitative agreement with the observations of Kri-voshaya (1983) and with the results of the model of Pinardi and Navarra (1993). In the Algerian Basin, a weak eastward coastal current develops.

6. General conclusions

Since the Mediterranean Sea was initialized with vertical profiles of temperature and salinity, its water mass
formation and circulation can easily be examined. In experiment E1, which includes all the forcings, the major observed features of the surface cyclonic circulation are adequately reproduced. As discussed in section 3, the transports through the Straits of Gibraltar and Sicily have values ranging within the observations, namely, 0.83 Sv for Gibraltar and 0.57 for Sicily. In the western Alboran Sea, the circulation vacillates between “an anticyclonic gyre” pattern and “a coastal current” pattern. In the Algerian Basin, the circulation displays large anticyclonic eddies. The surface transport is eastward as observed along the Algerian coast during the Mediprod IV cruise (Benzhora and Millot 1995a), but the model does not reproduce the eastward transport observed below 200 ms. The northern current is a permanent feature and shows realistic seasonal variations, with an increase in transport during winter. In the Liguro–Provencal Basin the cyclonic circulation is reinforced in winter, contributing to setting up the preconditioning phase for deep water convection. Significant water exchanges between the Liguro–Provencal and the Algerian Basins are found in the simulation. MAW flows northward between Sardinia and Menorca and along the west coast of Corsica more or less regularly. This transport could be associated with the Balearic front as observed on infrared satellite pictures by Deschamps et al. (1984) and to the current flowing along the Corsica coast found on current meter records by Taupier-Letage and Millot (1988). It is a significant source of freshwater (its salinity is less than 38 psu) for the Liguro–Provencal Basin and can contribute to inhibiting deep-water formation. It is driven by the atmospheric thermohaline forcing, as shown by the E’2 and E’2 sensitivity experiments. Besides, the surface basin-scale circulation displays a seasonal variability that modulates the exchanges between the sub-basins.
The advection of LIW seems too slow and too diffusive. This water mass fills up the Tyrrhenian Sea and is very slowly advected into the northern basin flowing along the western coast of Sardinia. Different reasons may be proposed to explain this slow circulation. First, the horizontal resolution is not high enough to give a proper representation of the advection of LIW, which flows as a coastal current with a characteristic length scale similar to a baroclinic radius of deformation of about 15 km. The inadequate resolution of topography, which is modeled by a $z$ scheme, may also explain the weak trapping of the current of LIW along the coast. Finally, the horizontal diffusion is obviously too high in the model: Instead of flowing along the Italian coast as it exits the Strait of Sicily, LIW spreads into the Tyrrhenian Sea.
Deep-water formation is inhibited after a few years in the model. This may be due to a too large advection of MAW in the MEDOC area as stated above or to a deficiency in the representation of the bottom topography trapping of the preconditioning phase, which plays an important role according to Hogg (1973) and Madec et al. (1996). Another deficiency might be introduced by the perpetual year forcing, which might not be representative of a typical year. The only way to overcome this difficulty would be to force the model with a decadal atmospheric forcing, which does not yet exist at this high spatial resolution. Besides, winter intermediate water is formed in different places in the basin: on the Iberian Shelf (this does not seem realistic because the discharge of the Ebro, which is not modeled, can prevent vertical convection), on the shelf of the Gulf of Lion and in the Balearic Sea, and in the center of the Liguro–Provençal Basin. Part of WIW flows southward into the Algerian Basin through the Ibiza Channel. This high amount of WIW may be a particular feature of the simulation in relation to the absence of deep-water formation.

Part I of this study (Herbaut et al. 1996) showed that the density gradients at the Straits of Gibraltar and Sicily are able to generate a basin-scale cyclonic surface circulation and contribute to 50% of the transport of the northern current. In the present study we found that the wind reinforces this circulation in opposition with Béthoux et al. (1982), who argued that the wind has no effect on the circulation of the Liguro–Provençal Basin. Idealized wind-driven experiments showed that this reinforcement can be interpreted by the generation of a planetary boundary layer current of Stommel’s type, while the cyclonic cell observed over the Ligurian basin is due to a Sverdrup balance associated with the wind curl. This cell participates to the preconditioning phase of deep-water formation, as argued in Madec et al. (1996), showing that the wind is an important ingredient of deep-water formation. Besides, these results show that the wind, and not the difference in the heat fluxes between the Tyrrhenian and the Ligurian Seas as suggested by Astraldi and Gasparini (1992), is responsible for transport variations in the Corsican Channel. The calculated transport and its fluctuations are in agreement with the observations of Astraldi and Gasparini (1992), this transport being a major momentum source for the Liguro–Provençal Current.

The above numerical experiments performed with a high-resolution numerical model help us understand the role of some mechanisms driving the circulation of the Western Mediterranean Sea. They clarify some hypotheses, such as the impact of wind versus thermohaline forcings, and indicate that further efforts must be devoted to the modeling of topographic effects, to increase the resolution of the model, and to get high-resolution decadal atmospheric forcings to avoid bias.

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\[\text{FIG. 22. A flat-bottom barotropic basin with an idealized geometry of the eastern part of the Western Mediterranean Sea is forced by a northwestern wind that is a simplification of the annual wind stress (the wind stress curl is linear, with a negative part in the southwestern part of the domain and a positive part in the northeast). The steady state is reached after 4 months of integration. In the northern part, a cyclonic circulation is associated with a coastal current along the northern coast.}\]

\[\text{FIG. 23. Experiment E3: Time evolution of (upper) the net volume transport through the Corsica Channel (in Sv, positive to the north) and (lower) the westward transport in the upper 200 m off Toulon (in Sv, positive to the west). The month on the abscissa indicates the end of the corresponding calendar month of the integration.}\]
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