Chapter 4

Changes in the Oceanography of the Mediterranean Sea and their Link to Climate Variability

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4.1. Introduction

In this chapter, we provide a description of the major features of the large-scale Mediterranean Sea circulation and the most important changes that have been observed in the physical oceanography of the basin. This chapter does not pretend to be a complete and comprehensive review of all the published works in the field but an introduction to the oceanography of the Mediterranean Sea aiming at an interdisciplinary audience of scientists working in various aspects of Mediterranean climate variability, the MedCLIVAR community. Most locations and topographic features mentioned in this chapter can be found in Fig. 2 of the Introduction to this book.

The basic circulation of the Mediterranean Sea has long been recognized to be that of a concentration basin (Marsigli, 1681; Waitz, 1755; Nielsen, 1912) and as such its major features are quite straightforward to describe. The excess evaporation over freshwater input within the basin (Garrett et al., 1993; Gillman and Garrett, 1994) is balanced by a two-layer exchange at the Strait of Gibraltar comprising a relatively warm, fresh (15°C, 36.2 psu) upper water inflow and a relatively cool and saltier (13.5°C and 38.4 psu) outflow to the Atlantic (Bryden et al., 1994; Tsimplis and Bryden, 2000).

The transformation of the inflowing Atlantic water to outflowing Mediterranean water is made through a thermohaline cell that involves the whole basin and leads to the formation of the Levantine Intermediate Water (LIW). In brief, the Atlantic inflow becomes progressively more saline as it moves eastwards. During winter it becomes cooler and denser and sinks to intermediate or sometimes deep levels. The role of the intermediate waters and in particular the LIW is important because they occupy the layer that corresponds to the sill depth (~280 m), thus determining the characteristics of the Mediterranean outflow. Waters deeper than the sill depth at the Strait of Gibraltar contribute only partly to the Mediterranean outflow (Stomell et al., 1973). In addition to the formation of intermediate waters, deep water formation takes place in several parts of the basin and in particular in the Gulf of Lions (Stommel, 1972; Mertens and Schott, 1998), the Adriatic Sea (Schlitzer et al., 1991; Malanotte-Rizzoli et al., 1997) and recently in the Aegean Sea (Roether et al., 1996). Therefore the Mediterranean Sea acts as a reduced scale ocean as regards the thermohaline circulation and dense water formation (Bethoux, 1980; Bethoux et al., 1998, 1999).

Existing climatological data sets indicate that the Mediterranean Sea is not in a steady state and is potentially very sensitive to changes in atmospheric forcing. In particular, during the last century several changes in the Mediterranean circulation have been documented. Trends in the temperature (T) and salinity (S) of the deep waters have been found in the Western Mediterranean (Lacombe et al., 1985; Charnock, 1989; Bethoux et al., 1990; Leaman and Schott, 1991; Rohling and Bryden, 1992; Bethoux and Gentili, 1999; Tsimplis and Baker, 2000) as well as the Eastern Basin (Tsimplis and Baker, 2000; Rixen et al., 2005) as well as in the upper waters (Painter and Tsimplis, 2003). The characteristics of the LIW have also been found to change in time (Astraldi et al., 1999; Brankart and Pinardi, 2001; Gasparini et al., 2005). Moreover sudden changes in the deep water formation sites in the Eastern Mediterranean have been documented between 1987 and 1995 (Roether et al., 1996).

This chapter discusses the oceanographic changes within the Mediterranean Basin concentrating mainly on the internal processes of the basin in relation to the regional forcing. Chapter 5 of this book (Artale et al.) integrates the Mediterranean Sea system to that of the North Atlantic oceanographic system and describes the characteristic mode of variability.

We start by describing the forcing of the Mediterranean Sea including the atmospheric forcing, the freshwater influx from rivers, the exchange with the Atlantic Ocean and the Black Sea (Section 4.2). Then the major features of the thermohaline circulation are described (Section 4.3) before we turn to the changes that have been detected in the basin (Section 4.4) and their links to atmospheric patterns.

We concentrate on the large-scale thermohaline circulation of the Mediterranean and we only examine sub-basin processes when these are directly linked with deep water formation. Thus we are not concerned with the complex combination of mesoscale and large-scale variations that dominate the surface Mediterranean circulation as seen from altimetry (Larnicol et al., 1995). However, we recognize that these also show significant variability and their contribution to the changes in the mean circulation is non-trivial (see for example, Millot, 1999).

In addition to the thermohaline circulation, we discuss two other key factors that are linked to the oceanic circulation, sea level variations (Section 4.6) and the wind-wave field (Section 4.7).

4.2. The Forcing of the Mediterranean Sea

We consider four forcing parameters of the Mediterranean Sea, namely, the air–sea interaction, the river influx, the exchange at the Strait of Gibraltar and the influence of the Black Sea.

4.2.1. Air–Sea Interaction

The circulation of the Mediterranean Sea is determined to a large extent by the air–sea exchanges of heat and freshwater, and the wind stress forcing of the basin. The exchange of salt and water at the Strait of Gibraltar (see Sections 4.2.3 and 5.3) provides a constraint on the estimates of the basin mean values of the heat and freshwater flux. Macdonald et al. (1994) obtain estimates of the equivalent basin mean net heat loss in the range 3–7 Wm^{-2} using mooring-based measurements of the transport. However, the available climatological estimates of the basin mean heat flux tend to show a net heat gain which is typically in the range 20–30 Wm^{-2} (Bunker et al., 1982; Garrett et al., 1993). This bias is presently believed to be caused by a combination of overestimated shortwave gain arising from inadequate parameterization of attenuation due to aerosols (Tragou and Lascaratos, 2003) and water vapour (Schiano, 1996) and underestimated longwave loss (Bignami et al., 1995).

Climatological annual mean fields based on the SOC climatology (Josey et al., 1999) for the net heat flux and the wind stress are shown in Fig. 73. The fields have been modified to include the Bignami et al. (1995) longwave formula as well as corrections to the shortwave formula of Gilman and Garrett (1994). A general north–south gradient in the net heat flux is apparent, from a net heat loss of up to 30 Wm^{-2} in the northern half of the basin to a gain of about 30 Wm^{-2} in the southern half. The gradient primarily reflects a reduction in the shortwave flux with increasing latitude and strong wind-driven latent heat loss in the Gulf of Lions, and the Adriatic and Aegean Seas. In winter, the heat loss in the latter three regions approaches 200 Wm^{-2} (e.g. Josey, 2003, Fig. 73) and is a major factor contributing to deep water formation. Significant interannual variations in the winter heat loss are known to occur, the prime example being the severe winters of the early 1990s which have been linked to the Eastern Mediterranean Transient (Theocharis et al., 1999b; Josey, 2003), discussed further in Section "when did the EMT start?". Broadly similar fields have been obtained in earlier studies (Bunker et al., 1982; Garrett et al., 1993).

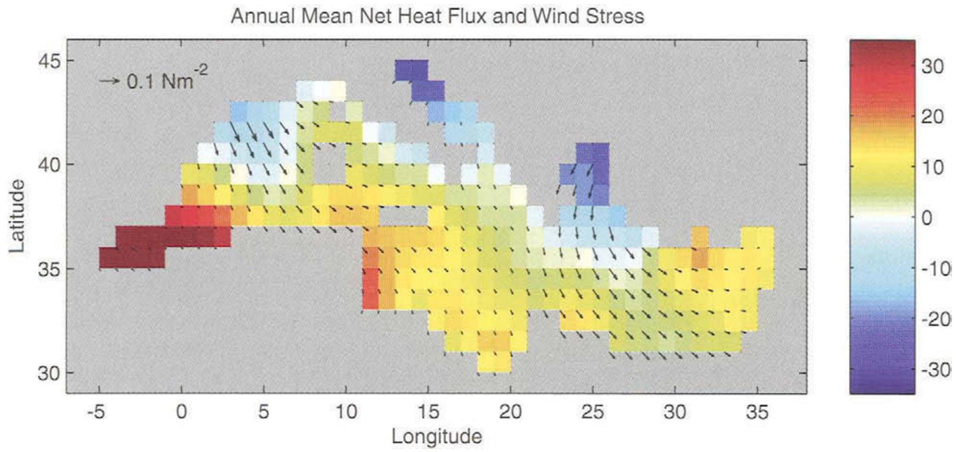


Figure 73: Climatological annual mean net heat flux (colours Wm^{-2}) and wind stress (arrows) from the modified version of the SOC climatology discussed in the text.

The basin mean E-P may be obtained via either the terrestrial or aerological branches of the hydrological cycle (Gilman and Garrett, 1994). In the terrestrial branch, the mean E-P is determined as the sum of river runoff into the basin and the freshwater flux through the Strait of Gibraltar, while for the aerological branch, the mean E-P is equated with the divergence of the vertically integrated horizontal water vapour flux. The corresponding estimates are E-P = 0.78 and 0.66 m per year from the terrestrial and aerological branches, respectively (Gilman and Garrett, 1994). The SOC flux climatology has a basin averaged of 0.71 m per year which lies between the two values noted above (Josey et al., 1999). The only region where the freshwater gains by precipitation and runoff exceed losses by evaporation is the Adriatic Sea (Raicich, 1996).

4.2.2. River Outflow

Published estimates of climatological values of river discharge in the Mediterranean Sea vary significantly (Table 4) due to different methods of evaluation and/or to the different datasets used.

Values for the discharge are in the range $8.1–16 \times 10^3\ m^3\ s^{-1}$ with the most recent estimates being lower than earlier estimates. Recently, Boukthir and Barnier (2000) obtain the river outflow to be $11 \times 10^3\ m^3\ s^{-1}$ while Struglia et al. (2004) obtain $8.1 \times 10^3\ m^3\ s^{-1}$ from the observations and $10.4 \times 10^3\ m^3\ s^{-1}$ by including an estimate of the errors associated with the unaccounted river basins. On the basis of the oceanic hydrological cycle, Pinardi et al. (2005) suggest that river runoff is in the range $2.2–9.8 \times 10^3\ m^3\ s^{-1}$, the upper end of this range

Table 4: Published estimates of river discharge in the Mediterranean Sea.

Author	**R ($10^3\ m^3\ s^{-1}$)**	**Method**
Tixeront (1970)	16.0	Rain maps, data, estimate of underground waters
Ovchinnikov (1974)	13.6	Observations
Margat (1992)	16.0	Total hydrological budget, including underground waters
Boukthir and Barnier (2000)	11.0	Observed river discharge from UNESCO dataset
Struglia et al. (2004)	8.1	Observed river discharge from Med-Hycos and GRDC datasets
Struglia et al. (2004)	10.4	As above, including correction to underestimates

is consistent with the lower values in Table 4. The discrepancy in mean annual river discharge between earlier and more recent studies may be due to the extensive damming of the rivers, to the fact that much more water has been used in the recent years for irrigation and to the impact of changes related to the North Atlantic Oscillation which have reduced precipitation and river runoff over the Mediterranean Sea.

An annual mean value for the river contribution in the Mediterranean Sea in the range $8.1–16 \times 10^3$ m^3 s^{-1} (Struglia et al., 2004) makes it less than 20% of the atmospheric freshwater flux (E-P) (Mariotti et al., 2002). Geographically, the dominant contributions to the runoff are from European rivers. The discharge into the Adriatic Sea, which is the major contributor, the Gulf of Lions, and the Aegean Sea together account for 62% of the total runoff. However, the contribution of the Middle Eastern rivers is likely to be heavily underestimated (up to 60%) because there are not sufficient published data for the Turkish rivers discharging in the Levantine Basin.

The North African discharge is mainly due to the Nile River which contributes about 1.4×10^3 m^3 s^{-1} (Struglia et al., 2004) or as low as 540 m^3 s^{-1} (Hamza et al., 2003) while contributions from other North African rivers are negligible (Struglia et al., 2004). The areas of maximum river discharge, namely the Adriatic, the Gulf of Lions and the Aegean Sea are also the areas where the strongest heat losses also occur in winter (Fig. 73) and are known deep water formation sites (see Section 4.3).

River runoff shows a seasonal cycle with an amplitude of about 5×10^3 m^3 s^{-1} the dry season being in midsummer and the peak flow in early spring (Fig. 74, from Struglia et al., 2004). The amplitude of the seasonal cycle of river outflow is almost negligible when compared to the seasonal cycle of E-P (Boukthir and Barnier, 2000; Mariotti et al., 2002). However, because the phase of the seasonal cycles of E-P and river outflow differ, the two basin-integrated components E-P and R are comparable during spring. Thus, on a sub-basin scale, river discharge variability may be responsible for some modulation of the oceanic processes (Rohling and Bryden, 1992; Zavatarelli et al., 1998; Bethoux and Gentili, 1999; Boscolo and Bryden, 2001). However, late spring freshwater discharges are unlikely to directly affect deep water formation which may probably occur earlier during the year.

Observed interannual variations in runoff during the twentieth century have been up to 60% of the climatological long-term mean, while decadal variations are of the order of 20% of the long-term mean (Struglia, 2004). The interannual and/or decadal variability of river runoff is expected to be dominated by the major factor controlling precipitation over Europe, that is, the North Atlantic Oscillation (NAO). Northern European river flows are positively correlated, particularly in winter but also in spring, with the NAO Index while rivers located

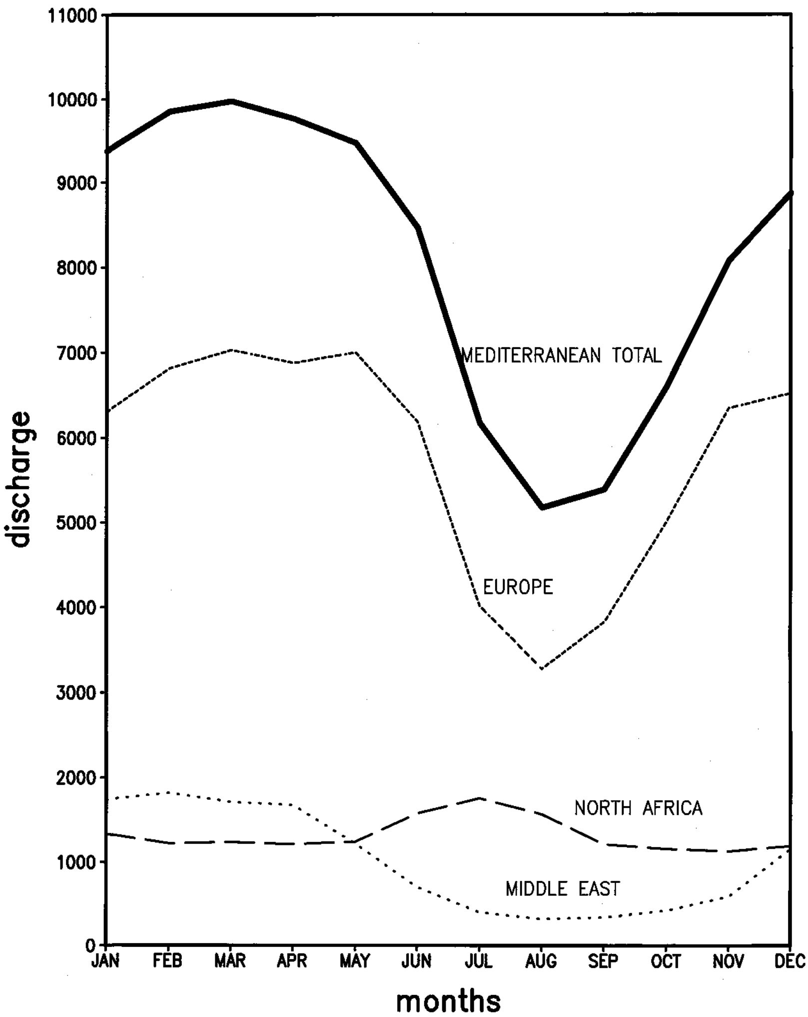
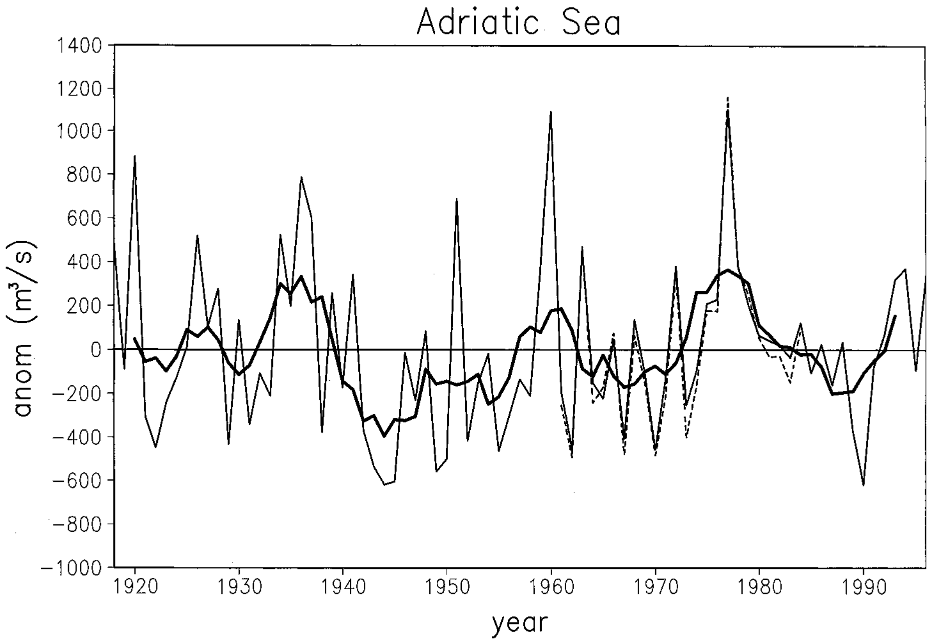


Figure 74: Climatological seasonal cycle of total discharge into the Mediterranean Sea and its decomposition by continent of origin (values are in $m^3 s^{-1}$). From Struglia et al., 2004.

in South Europe are negatively correlated (Shorthouse and Arnell, 1997). The interannual variability related to the NAO pattern has been confirmed in various Mediterranean rivers (Send et al., 1999; Cullen et al., 2002; Struglia et al., 2004; Trigo et al., 2004). The annual mean anomalies for Po, in the Adriatic, and the Rhone, in the Gulf of Lions, are shown in Fig. 75.

The effects of other global patterns on the Mediterranean river discharge have not yet been resolved fully. However, ENSO has been suggested as affecting the Nile stream flow (Elfatih and Eltahir, 1996; De Putter et al., 1998) while



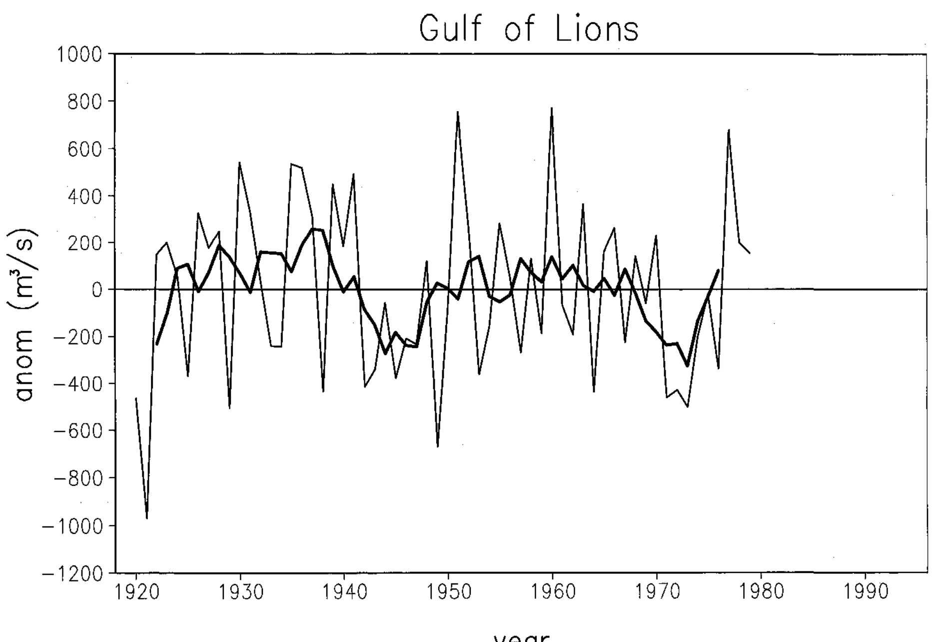


Figure 75: Anomalous river discharge in the Adriatic Sea (top) and in the Gulf of Lions (bottom). A reconstructed time series for the Adriatic is shown for the period 1961–84 (dashed line). Over the period 1918–96 the Adriatic time series is represented by that of the Po river (thin solid line). Discharge in the Gulf of Lions is that from the Rhone river for the period 1920–79. In both panels thick solid lines are the 5-year running means.

mid-latitude stream flow responses to extreme phases of Southern Oscillation have been detected over Turkey (Kahya and Karabörk, 2001).

4.2.3. Exchanges through the Strait of Gibraltar

The exchanges of the Mediterranean Sea both at the Strait of Gibraltar and at the Dardanelles are two-way flows with salinity differences driven by the water mass characteristics of the basins and mass exchanges limited by the hydraulic control points of the Straits. The exchange at the Strait of Gibraltar is of paramount importance for the Mediterranean Sea. However, as in this book it is dealt with under Section 5.3 (by Artale et al., in this book) we will deal with it briefly. Candela et al. (1989) have explained up to 80% of the observed variability of the currents in the strait on the basis of an almost barotropic mode. The inflow of Atlantic water is highly modulated by the tidal signal and the atmospheric forcing at the Strait of Gibraltar. Its barotropic component has been found to correlate well with the cross-strait sea level component (see for example, Tsimplis and Bryden, 2000). Estimates of the Atlantic inflow spanning much of the last century range between 0.72 and 1.75 Sv while outflow values range between 0.67 and 1.68 Sv (Table 5). It is not clear to what extent the variability in the estimates is a reflection of changes in the freshwater balance or an artefact of the technique used for the estimates. However omitting the oldest estimate of Scott (1915) significantly reduces the range of the inflow between 0.72 and 1.26 Sv. The net exchange is of the order of tenths of Sv ($1\ \text{Sv} = 10^6\ \text{m}^3\ \text{s}^{-1}$).

The physics of the exchange are further discussed in Section 5.3 of this book.

Table 5: Published estimates of the exchange at the Strait of Gibraltar.

Authors	Inflow (Sv)	Outflow (Sv)	E-P-R (Sv)
Schott (1915)	1.75	1.68	0.07
Bethoux (1979)	1.26	1.20	0.06
Perkins et al. (1990)	1.3–1.6		
Bryden and Kinder (1991)		0.80	
Bryden et al. (1994)	0.72	0.68	0.04
Hopkins (1999)	1.26	0.84	0.42
Tsimplis and Bryden (2000)	0.78	0.67	0.10
Baschek et al. (2001)	0.81	0.76	0.05
La Fuente et al. (2002)	0.97	0.84	0.13

4.2.4. The Exchange with the Black Sea

The basic dynamical feature in the Black Sea is the cyclonic rim current encompassing the sea approximately along the 1,000 m isobath and caused by the combined impacts of both mechanical and thermohaline forcing (Stanev, 1990; Oguz and Malanotte-Rizzoli, 1996; Staneva et al., 2001; Beckers et al., 2002; Kara et al., 2005). In addition, numerous mesoscale eddies created by the main flow instabilities are observed near the coast (Fig. 76). The general circulation of the BS is subject to pronounced seasonal variations as it is mainly wind-driven.

Another permanent major feature of the Black Sea Basin is the Cold Intermediate Layer (CIL), which is due to the existence of a strong halocline at 70–100 m depth which limits the winter convection. The replenishment time of the CIL is estimated to be ~5 years (Lee et al., 2002). As seen from Fig. 77 the amount of Cold Intermediate Water within the CIL can be substantially reduced after several warm winters thus demonstrating the sensitivity of the basin to climate variability.

The Mediterranean water that enters the Black Sea sinks below the CIL as a salt wedge of 10–20 m near the Bosporus, decreasing rapidly on the shelf (<5 m) and becoming almost indistinguishable along the continental slope because of the large entrainment and dilution. The sinking is channelled by the

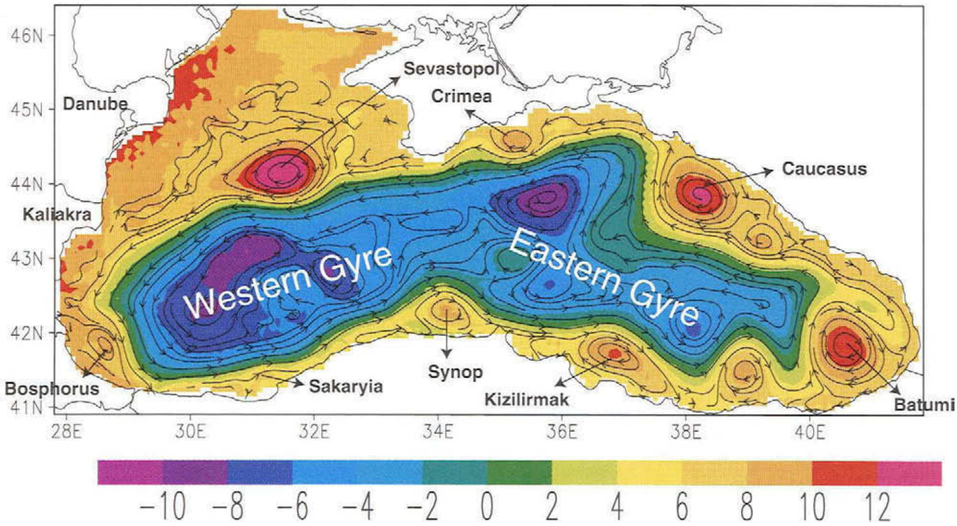


Figure 76: Snapshot of sea level (cm) and surface streamlines simulated by the DieCAST model (see for more details, Staneva et al., 2001).

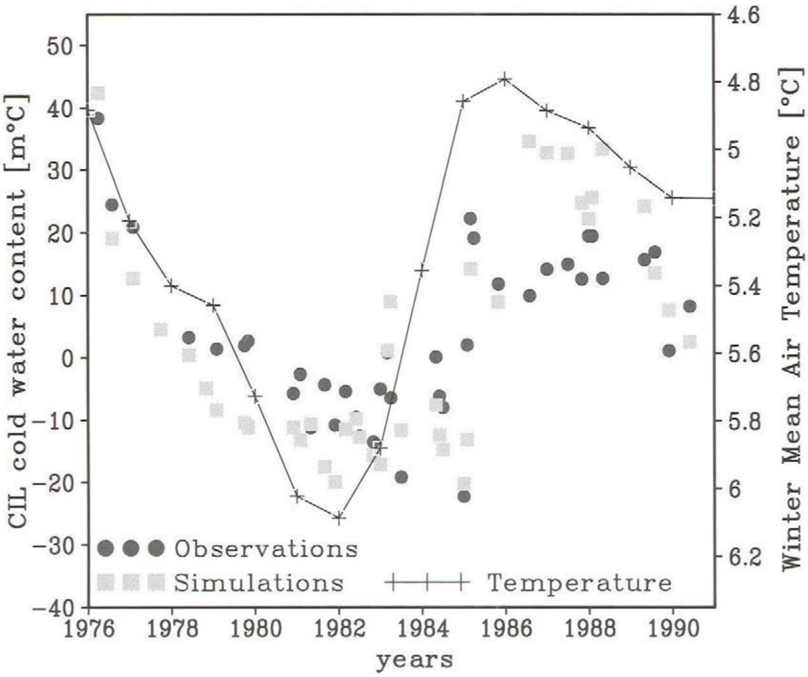


Figure 77: Temporal variability of the cold water content between isopycnal levels 14.4 and 15.6 sigma. Observations and simulations with the Black Sea Modular Ocean Model (Stanev et al., 2004) are shown. The solid line (right *y* axis) gives the winter (DJF) mean air temperature at surface estimated from ERA40 reanalysis data set.

underwater extension of the Bosporus Strait and propagates further on the shallow shelf, sinking abruptly at the continental slope (Yuce, 1996; Gregg and Ozsoy, 1999; Stanev et al., 2001). Analyses of observations and numerical modelling suggest that the penetration of saline Mediterranean waters is limited to the upper 400–600 m and does not reach the bottom of the Black Sea. However, the inflowing Mediterranean water is the origin of massive intrusions of oxygen-enriched water extending to ~200 km from the coast thus contributing to the lateral ventilation of the Black Sea (Konovalov et al., 2003).

The inflow of water from the Black Sea to the Mediterranean is about two orders of magnitude smaller than the inflow of Atlantic water through the Strait of Gibraltar. However, the salinity difference between the Black Sea and Mediterranean Sea water is of ~18 psu, that is, much larger than the ~2 psu differences in the Strait of Gibraltar. Because of the large salinity differences, the role of the Black Sea outflow is not negligible, at least for the Aegean Sea. The salinity contrast between the Mediterranean and the Black Sea triggered the first ever attribution of marine circulation to density differences (Marsigli, 1681). At the Bosphorus Strait (sill depth ~36 m) the Black Sea water occupies the surface layer, while the Mediterranean Sea water is observed as a salt wedge in the bottom layer. The mean position of the interface which is ~10 m thick is well represented by the 20 psu isohaline. The thickness of these layers as well as their transports, are very sensitive to the sea-level difference between the basins which in turn is controlled by the basin's water balance, regional circulation intensity and direction of wind. The two-layer density stratification separated by a transition layer is almost always observed in the Bosporus and Dardanelles Straits. However, blocking of the lower layer flow occurs during combinations of high water in the Black Sea and northerly winds (Latif et al., 1991; Yuce, 1996), which are short lived. Similarly blocking of the surface current occurs during southerly winds (Gunnerson and Ozturgut, 1974). Mixing and turbulent entrainment processes in the Straits dominate the water, salt and heat transport between the original reservoirs (Besiktepe et al., 1994; Gregg et al., 1999; Stanev et al., 2001). The mean vertically integrated transport in the Dardanelles and Bosphorus Straits on the basis of the water budget and sea level variation was estimated as 6×10^3 m^3 s^{-1} (Stanev and Peneva, 2002).

Recent analyses of hydro-meteorological and oceanographic data demonstrate a strong correlation between the North Atlantic Oscillation, sea level variability and thermal state of the Black Sea (Stanev et al., 2000; Stanev and Peneva, 2002; Oguz et al., 2003, Tsimplis et al., 2004).

The inflowing in the Aegean Sea, Black Sea water occupies the surface layers in the North Aegean Sea where it is thought to have a controlling function on the vertical stability and mixing (Zervakis, 2000; Stanev and Peneva, 2002).

Variations of the Black Sea water outflow may affect the thermohaline circulation in the North Aegean Sea (see Section "Why did the EMT happen?"). Reductions of ~100 km^3/year are quite plausible, which are equivalent to changes in evaporation of 0.2 m/year over the Aegean Sea (Stanev and Peneva, 2002). By contrast, an increase of the transport of Black Sea water into the Mediterranean Sea could block or at least decrease the rates of any deep water formation taking place in the North Aegean Sea.

The sea level in the Black Sea is the most important parameter controlling the exchange between the basins and it is subject to large seasonal, interannual and decadal variability.

The interannual and decadal sea level variability is mainly a response to the external hydrological forcing (Stanev and Peneva, 2002; Tsimplis et al., 2004). The sea level anomalies in the time series of Batumi and Sevastopol (Fig. 78) show variations clearly related to the fresh water flux calculated as the net sum of river runoff, precipitation and evaporation over the whole Black Sea area (Simonov and Altman, 1991).

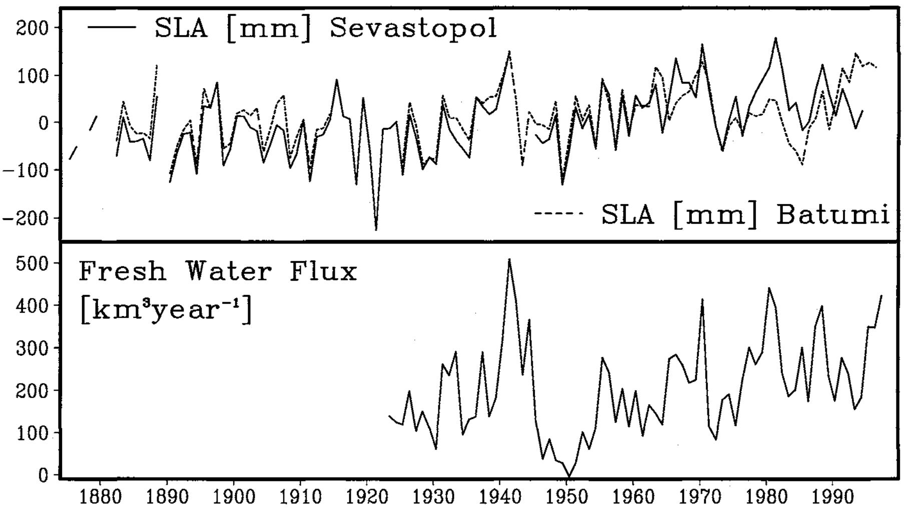


Figure 78: Annual mean sea level anomalies (mm) in Batumi and Sevastopol on the upper panel. The lower panel shows the annual mean values of fresh water flux, including river runoff, precipitation and evaporation (km^3/year).

For the period 1923–1997, the mean values of evaporation (E) and precipitation (P) are respectively 12×10^3 m^3 s^{-1} and 7.6×10^3 m^3 s^{-1} (Simonov and Altman, 1991; Stanev and Peneva, 2002). Thus evaporation exceeds precipitation but the freshwater balance becomes positive because of the river runoff.

The annual discharge of the six (biggest) rivers, Danube, Dniepr, Dniestr, Southern Bug, Don and Kuban, in the Black Sea is 8.6×10^3 m^3 s^{-1} (270.3 km^3/year) (Fekete et al., 1999), that is, equivalent to the river outflow in the whole of the Mediterranean Sea. There are large uncertainties in the above-stated estimates (Simmons and Gibson, 2000). Thus, further research is needed in order to understand the regional water balance and its sensitivity to climate change (Hagemann et al., 2005).

4.3. The Mediterranean Overturning Circulation

4.3.1. The Modification of the Atlantic Inflow

A simplified sketch of the Mediterranean Overturning Circulation is provided in Fig. 79. The main supplier of fresh water for the Mediterranean is the Atlantic Ocean, at an average rate of about 0.8 Sv. The Atlantic inflow enters as a surface current of mainly Atlantic characteristics (S ~ 36.5), slightly modified through mixing with the outflowing Mediterranean waters. The Modified Atlantic Water (MAW) moves towards the east following the mean cyclonic circulation

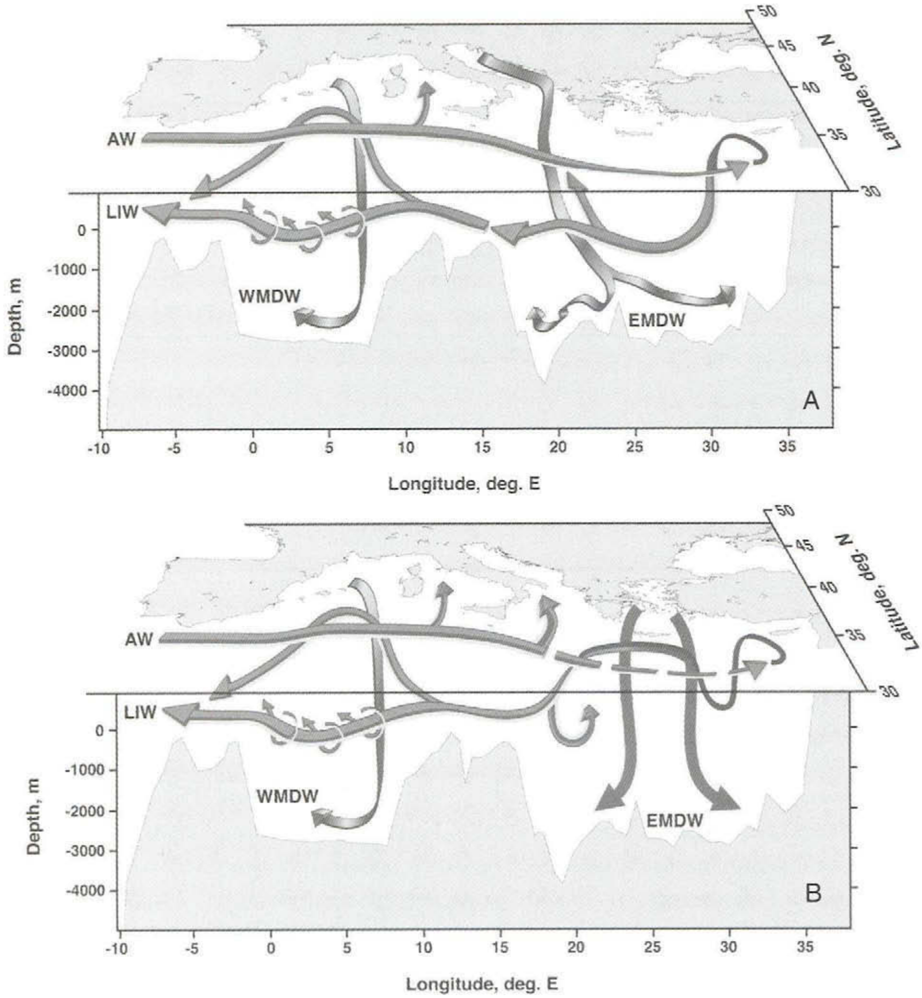


Figure 79: Sketches of the Mediterranean Overturning Circulation before the EMT (A) and during the EMT(B). Before the EMT, the MAW moved eastwards and forms about one-third of its transport as the Levantine Basin (LIW) which is then tranported westwards branching to the Adriatic and crossing back to the western basin. After crossing the Strait of Sicily it follows the coasts of the Western Mediterranean anticlockwise. There is a disagreement on whether there is a branch crossing the western basin westwards or whether large eddies (symbolized by the successive circles) are responsible for the westward transport. Deep water formation of about 0.3 Sv took place in the Adriatic Sea and the Gulf of Lions. During the EMT (B) the MAW is deflected northwards into the north of the Ionian Basin and its eastward tranport diminishes. The intermediate water formation is then deflected northwards into the Cretan Basin where about 1 Sv of deep water is formed. In this sketch, we have kept the Western Mediterranean the same as before the EMT. However evidence of impacts of the EMT in the Western Mediterranean have started being published (see Section 4.4.3 for details).

of the basin and, because of the increased warming and evaporation its salinity increases gradually to about 37.0–37.5 in the Strait of Sicily (Astraldi et al., 2002) and to around 38.6 in the vicinity of the Cretan Passages (Malanotte-Rizzoli et al., 1997). By contrast the depth of the salinity minimum increases eastward (from 20 to 100 m) during summer and autumn. However not all MAW crosses to the Eastern Mediterranean but some stays in the western basin which acts as a pool of MAW (Millot, 1999). The amount of MAW crossing to the Eastern Mediterranean is not known. However Millot (1999) indicates values around 1–3 Sv.

The flow of MAW from the Strait of Gibraltar to the Strait of Sicily and then to the Eastern Mediterranean is part of the basic cyclonic circulation of the Western Mediterranean Basin which essentially is comprised of an anti-clockwise cell complicated by the islands of Corsica and Sardinia. The MAW follows the North African coasts moving eastwards and forming the unstable Algerian current which generates both short-lived mesoscale eddies inconsequential for the basin-wide circulation and larger scale open sea eddies which detach from the main current and propagate seawards before turning westwards thus mixing MAW with the interior of the Western Mediterranean Basin (Millot et al., 1997; Fuda et al., 2000). At the Strait of Sicily the MAW separates into two branches, one which crosses to the Eastern Mediterranean while the other branch moves northward into the Tyrrhenian Sea flowing anticlockwise along the topography of the Italian coasts of the Western Mediterranean. At the most northern point of the Tyrrhenian Sea a further separation occurs. One branch goes north of Corsica and joins the westward flow of the second branch which first completes the circle around the Tyrrhenian Sea and then moves northwards along the Corsican and the Sardinian coasts. The rejoined branches form the Northern Current (Millot, 1999). The Northern Current continues along the French and Spanish coasts and separates into two branches one moving into the south of the Balearic Sea (Millot, 1999), while the second part continues southwards and can be found at depths greater than 200 m in the Alboran Sea as old MAW that has completed one or possibly more rounds of the Western Basin (Millot, 1999). During its circulation in the Western Mediterranean Basin the salinity of the MAW increases to 38.0–38.3 near the north coasts of the Western Mediterranean (Millot, 1999). For comparison the surface salinity at the vicinity of the Cretan passage is about 38.6 psu.

4.3.2. Intermediate Water Formation

The high salinity preconditions the surface waters for intermediate or deep water formation. In the Levantine region, surface waters reach salinities of ~38.9 psu.

The density of the surface waters increases further through evaporation and cooling during winter until it is dense enough to sink. Depending on where the sinking water will settle, either Levantine Intermediate Water (LIW) or Levantine Deep Water (LDW) is formed. The LIW is the main water-mass of the Mediterranean Sea, occupying the intermediate layers between 200 and 500 m. Its core is usually identified by a salinity maximum of 38.95–39.05 psu near the production areas (Lascaratos, 1993) which reduces to 38.75 psu at the Strait of Sicily (Bethoux and Gentili, 1996, 1999; Astraldi et al., 1999, 2002) and to 38.4 psu at the Mediterranean outflow at the Strait of Gibraltar (Tsimplis and Bryden, 2000). The amount of LIW formed annually ranges in model outputs between 0.6 and 1.3 Sv with a typical climatological value of 1.0 Sv (Nittis and Lascaratos, 1998) which is consistent with previous estimates, based on a variety of different methods (Ovchinnikov, 1984; Lascaratos, 1993; Tziperman and Speer, 1994).

One region in which either LIW or LDW has been observed to be produced is the Rhodes gyre (Ovchinnikov, 1984; Ovchinnikov and Plakhin, 1984; Malanotte-Rizzoli and Hecht, 1988; Buongiorno Nardelli and Salusti, 2000). The development of the cyclonic Rhodes gyre is crucial in the production of the LIW because of the doming of the isopycnals at the centre of the gyre (Lascaratos et al., 1993; Lascaratos and Nittis, 1998) which reduces the surface layer and brings the subsurface denser water masses closer to the influence of the surface atmospheric forcing. Modelling studies indicate that the LIW formation area can either expand over the whole of the North Levantine Basin or be shifted southwards either by changes in the mean buoyancy loss or due to synoptic-scale forcing (Nittis and Lascaratos, 1998). Other sites that have been suggested as candidates for LIW formation are the south Aegean Sea and the south Levantine Sea (Wüst, 1961; Bruce and Charnock, 1965; Morcos, 1972; Ozturgut, 1976; Theocharis et al., 1988; Georgopoulos et al., 1989).

Intermediate waters of maximum salinity around 38.3 psu are also formed during winter in the northern coasts of the Western Mediterranean (Millot et al., 1999 and references therein). These are found underlying the MAW. They are termed the Winter Intermediate Water. Millot (1999) considers the process of formation of WIW as important in transforming the MAW into Mediterranean water types, although the exact role of the WIW is not yet resolved. The WIW follows the general circulation of the western basin underneath the MAW.

The path of the LIW is normally westwards south of Crete and then filling in the intermediate waters of the Ionian Sea branching through the Strait of Otranto into the Adriatic while another branch moves towards the west possibly following the east coasts of Italy southwards and then crossing to the western basin. After it crosses the Strait of Sicily westwards, the LIW, mixes quickly and modifies the waters from ~200 to ~1800 m (Millot, 1999). The intermediate waters in the western basin appear to follow a similar-to-the MAW anticlockwise pattern around the Tyrrhenian Sea and then northwards along the west coasts of Sardinia and Corsica and underneath the Northern current following the topography (Millot, 1999).

However, there is significant debate on the return flow of the LIW in the Western Mediterranean. One issue is whether circulation of the LIW around the Tyrrhenian Sea actually occurs or whether the LIW crosses directly to the southern tip of Sardinia (see for example, Roussenov et al., 1995). The second issue is whether the flow of the LIW towards the Strait of Gibraltar does indeed take place following the anticlockwise circulation path suggested by Millot (1987, 1999) or whether a direct westwards flow across the western basin exists as several models indicate (Wu and Haines, 1996; Herbaut et al., 1997). A recent study (Millot and Taupier-Letage, 2005) reinforces the view of a full anticlockwise circulation and explains the LIW found in the central Algerian Basin as a result of the transport by mesoscale eddies detached from the Algerian current rather than a continuous westward flow (Millot and Taupier-Letage, 2005). The disagreement between the observational analyses (Millot, 1987, 1999; Millot and Taupier-Letage, 2005) and the models which indicate the existence of a westward flow is puzzling. The authors are inclined to accept the suggestion that model deficiencies are the major cause of the discrepancy, however as the most recent data were collected in the 1990s, which is not a typical decade for the circulation of the Mediterranean due to the EMT and the associated changes in all levels of the Eastern Mediterranean, some effects in the spreading of the intermediate waters in the western basin cannot be excluded. Indeed Gasparini et al. (2005) observe increases in the density in the Sicily westward flow and a deep cascade in the Tyrrhenian deep water thus explaining the deep water trends observed by Fuda et al. (2002), Astraldi et al. (2002) in the Tyrrhenian Sea. The deeper than usual cascade of the incoming LIW could also cause an interruption of an intermediate westward flow across the western basin, although some of it would be trapped by the deep eddies detached from the Algerian current as observed by Millot and Taupier-Letage, 2005.

4.3.3. Deep Water Formation in the Eastern Mediterranean

Deep water formation takes place both in the Eastern and the Western Mediterranean Basins. In the Eastern Basin, surface waters from the Levantine that have not participated in the formation of LIW continue their circulation turning northwards and westwards, to reach the northern coasts of the Aegean and Adriatic Seas. During winter under the influence of cold, dry northerly winds they become colder and even more saltier and depending on the strength of the heat losses may sink thus producing deep waters.

Until recently, the Adriatic Sea was the only generally accepted area of deep water formation (Schlitzer et al., 1991; Malanotte-Rizzoli et al., 1997). The salinity of the Eastern Mediterranean Deep Water mass has always been found less than the Aegean waters outflowing the Cretan Arc Straits, thus its origin has undoubtedly been the Adriatic Deep Water (Pollak, 1951; Wüst, 1961; Hopkins, 1978, 1985; Schlitzer et al., 1991). These views were to change completely in 1996.

The general circulation over the Southern Adriatic is permanently cyclonic and controlled by bottom topography (Manca and Giorgetti, 1998; Kovacevijć et al., 1999; Poulain, 2001). The intrusion of LIW through Otranto Strait (Buljan and Zore-Armanda, 1976) is a process enhancing the potential for dense water production in the Southern Adriatic (Manca et al., 2002). Winter conditions favour the intensification of the cyclonic circulation and the doming of the isopycnals in the centre of the gyre, thus facilitating open-sea convection in the Southern Adriatic. Dense water formation takes place both in the shallow Northern Adriatic and the deep southern Adriatic (Ovchinnikov et al., 1987; Manca et al., 2002). The convection process over the northern shelf has the characteristics of shelf formation, under the influence of cold and dry northerlies in the winter (Franco et al., 1982; Malanotte-Rizzoli, 1991), while the formation over the deep Southern Adriatic has the character of open ocean convection (Manca et al., 2002).

The dense waters produced over the Northern Adriatic by shelf formation propagate to the south and mix with the dense water produced in the deep southern Adriatic, to form the Adriatic Deep Water (ADW) (Manca et al., 2002), which eventually overflows as an undercurrent through the Otranto Strait and constitutes the main contributor to the Eastern Mediterranean Deep Water (EMDW). The North Adriatic dense waters are characterized by densities $\sim$29.4–29.9 kg m^{-3} (Franco et al., 1982; Malanotte-Rizzoli, 1991). The annual export of ADW towards the Ionian Sea, measured throughout the period 1997–1999, ranged between 0.1 Sv in 1997 and 0.4 Sv in 1999, with fluctuations reaching 1 Sv (Manca et al., 2002). For the period 1985–1987, it was estimated that the renewal time for the EMDW waters formed in the Adriatic is about 126 years (Roether and Schlitzer, 1991; Schlitzer et al., 1991; Roether et al., 1994).

The maximum salinity of the Adriatic appears to scale well with an index defined as the atmospheric pressure difference between the central Mediterranean and the mid-Northern Atlantic (Grbec et al., 2003). Some similarity with the low-frequency NAO variation was also evident in the same study and in Tsimplis and Rixen (2002).

The Aegean sub-basin was recognized as a potential contributor to the waters filling the deep Ionian and Levantine Basins, the Eastern Mediterranean Deep Water (EMDW) by several researchers (Nielsen, 1912; Lacombe and Tchernia, 1960). However, although dense Aegean waters have been recorded exiting the Straits of the Cretan Arc (Nielsen, 1912; Lacombe and Tchernia, 1958; Miller, 1974; El-Gindy and El-Din, 1986), their amount and density were not considered enough to constitute a significant contributor to the EMDW waters (Schott, 1915; Pollak, 1951; Wüst, 1961; Hopkins, 1978, 1985).

It is only recently that the role of the Aegean Sea as a deep water formation area has been conclusively demonstrated (Roether et al., 1996). It is now known that the Levantine Surface Water can reach surface salinity which exceeds 39.0 in the Aegean Sea (Zervakis and Georgopoulos, 2002) and bottom water formation has been recorded over the Samothraki and Lemnos plateaux (Ovchinnikov et al., 1990; Theocharis and Georgopoulos, 1993) while the highest density ever recorded in the North Aegean is 29.64 kg m^{-3} (Zervakis et al., 2000). The shallow Cyclades plateau has also been suggested as a potential source of very dense waters (Theocharis et al., 1999a).

Curiously the recently observed initiation of dense water formation in the Aegean Sea was accompanied by diminution of deep water formation in the Adriatic. This would imply that the production site depends on the supply of LSW and possibly LIW in order to produce deep waters and that the supply is mainly to one or the other region.

4.3.4. Deep Water Formation in the Western Mediterranean

In the Western Mediterranean, deep water formation takes place in the Gulf of Lions (Stommel, 1972) where the heat-dominated local buoyancy flux appears to determine the depth of the deep water convection (Mertens and Schott, 1998). During winter, dry and cold air initially mixes the MAW and the WIW with the underlying, warmer and saltier LIW. Further heat loss leads to formation of Western Mediterranean Deep Water (WMDW). The formation of the WMDW depends on a preconditioning period, followed by violent mixing lasting for about an hour and in convection cells about 1 km in diameter. Finally, the deep water formed spreads out of the convective region (Reihn, 1995). The deep water formation appears to be a multiscale process with small-scale 1 km cells occurring within a region of about 100 km and spreading out as instability eddies of scale 5–10 km (Reihn, 1995). The rate of formation from the same event has been estimated to be between 0.2 and 0.3 Sv (Send et al., 1995, 1999) to around 3 Sv (Reihn, 1995) depending on the method used. Deep water formation does not always reach the full depth of the basin although in most years it does (Send et al., 1999). In less severe winters the waters settle at intermediate depths. Send et al. (1999) suggest that variable deep water formation driven by varying local atmospheric forcing linked to NAO-related variability is responsible for the observed changes in the deep water characteristics (see the discussion in Section 4.4.1). Significant interannual variability in the amount of the deep waters formed as well as the location is suggested by Millot (1999). In addition to the open sea formation, dense water formation may also take place on the shelf (see for example, Durrieu de Madron et al., 2005). Fuda et al. (2002) have recently suggested that deep water is also formed in the Tyrrhenian Sea although it appears that this is now resolved by Gasparini et al. (2005) as waters coming from the Eastern Mediterranean.

4.4. Climatic Changes in the Mediterranean Sea Circulation

4.4.1. Multidecadal Trends in Water Mass Characteristics

Deep water changes have been observed both in the Eastern and the Western Mediterranean Basin. Some of these changes are linked, others may be independent as the sills at the Strait of Sicily restrict the exchange of deeper waters between the basins.

The deep water temperature and salinity in the western basin have been found to show interannual, decadal and longer term variability. Changes in the Western Mediterranean Deep Water (WMDW) have been identified by several authors (Lacombe et al., 1985; Charnock, 1989; Bethoux et al., 1998). The WMDW has become warmer and saltier over the period 1959–1997 (Bethoux et al., 1998) with trends of 3.5×10^{-3}°C year^{-1} and 1.1×10^{-3} year^{-1} for the deeper layers and trends of 6.8×10^{-3}°C year^{-1} and 1.8×10^{-3} year^{-1} for the intermediate layers (Bethoux and Gentili, 1999). Various causes for these trends, including anthropogenic influence, local atmospheric conditions and hydrological conditions during dense water formation events, as well as the first signature of global warming have been proposed. Bethoux and Gentili (1999) argue that while the warming trend can be explained by greenhouse-effect-related warming of the sea surface between 1940 and 1995, the salinity trends require a rate of increase in the water deficit of the Mediterranean of the order of 0.10 m year^{-1}. In order to achieve such a high value, it is necessary to consider not only the damming of the major Nile (Wadie, 1984) and Ebro (Martin and Milliman, 1997) rivers, but also a small increase in evaporation and decrease of the net Black Sea outflow, mainly due to Central/Eastern European river damming (Tolmazin, 1985; Bethoux and Gentili, 1996). A model study

(Skliris and Lascaratos, 2004), suggests that the river Nile damming could account for about 45% of the salinity increase in the WMDW.

In examining the causes of the temperature and salinity trends of the Western Mediterranean, Krahman and Schott (1998) identified increasing trends in salinity in the surface and bottom layers, and partly attributed these trends to a gradual drop in precipitation associated with an increasing trend of the index of the North Atlantic Oscillation (Hurrell, 1995) from 1960 until 1990, as well as a decrease of the Ebro river outflow (Ibanez et al., 1996; Martin and Milliman, 1997; Tsimplis and Josey, 2001). Furthermore, Krahman and Schott (1998) related the deep water changes to local conditions in the WMDW formation area in the Northwestern Mediterranean.

Estimates of trends in the Eastern Mediterranean have been included in several analyses (Tsimplis and Baker, 2000; Painter and Tsimplis, 2003; Manca et al., 2004). These analyses were made possible by the coordinated efforts that led to the development of much needed regional oceanographic databases, initially the Mediterranean Oceanographic Data Base (MODB) (Brasseur, 1995; Brassuer et al., 1996) and subsequently the MEDATLAS (Maillard et al., 2001; MEDATLAS group, 1997). Further progress was made by the inclusion of ex-USSR data (Hecht and Gertman, 2001) which cover areas in the Gulf of Libya where previously large gaps existed. Tsimplis and Baker (2000) identified significant increasing trends in the temperatures of the deep waters of the Eastern Mediterranean and some suggestions for trends in salinity, though they were reluctant to accept them as real, due to the large scattering of the salinity values especially in the Ionian Sea. Painter and Tsimplis (2003) have found that the upper waters of the entire Eastern Mediterranean have been undergoing a sustained period of cooling throughout all seasons since about 1950. The identified trends were caused by significant reductions in the winter temperatures while the other seasons did not have in general statistically significant signals. The salinity was also found to be increasing over the same period with the strongest trends often being found at the shallowest horizontal levels examined. The upper layers of the Western Mediterranean exhibit the same salinification as seen in the Eastern Mediterranean but the temperature trend is restricted to the Eastern Mediterranean. The analysis of Painter and Tsimplis (2003) supports the suggestion made by Krahmann and Schott (1998) that the LIW is not the major contributor to changes in the deep water characteristics of the Western Mediterranean implying that changes in the local atmospheric forcing are responsible for the deep water trends. However, Manca et al. (2004), have identified trends in the deep waters of the Eastern Mediterranean (see Figs. 80, 81) of similar order of magnitude as those estimated in the western basin by Bethoux et al. (1990) implying either a common atmospheric origin or a communication of the changes from one basin to the other most likely through the LIW.

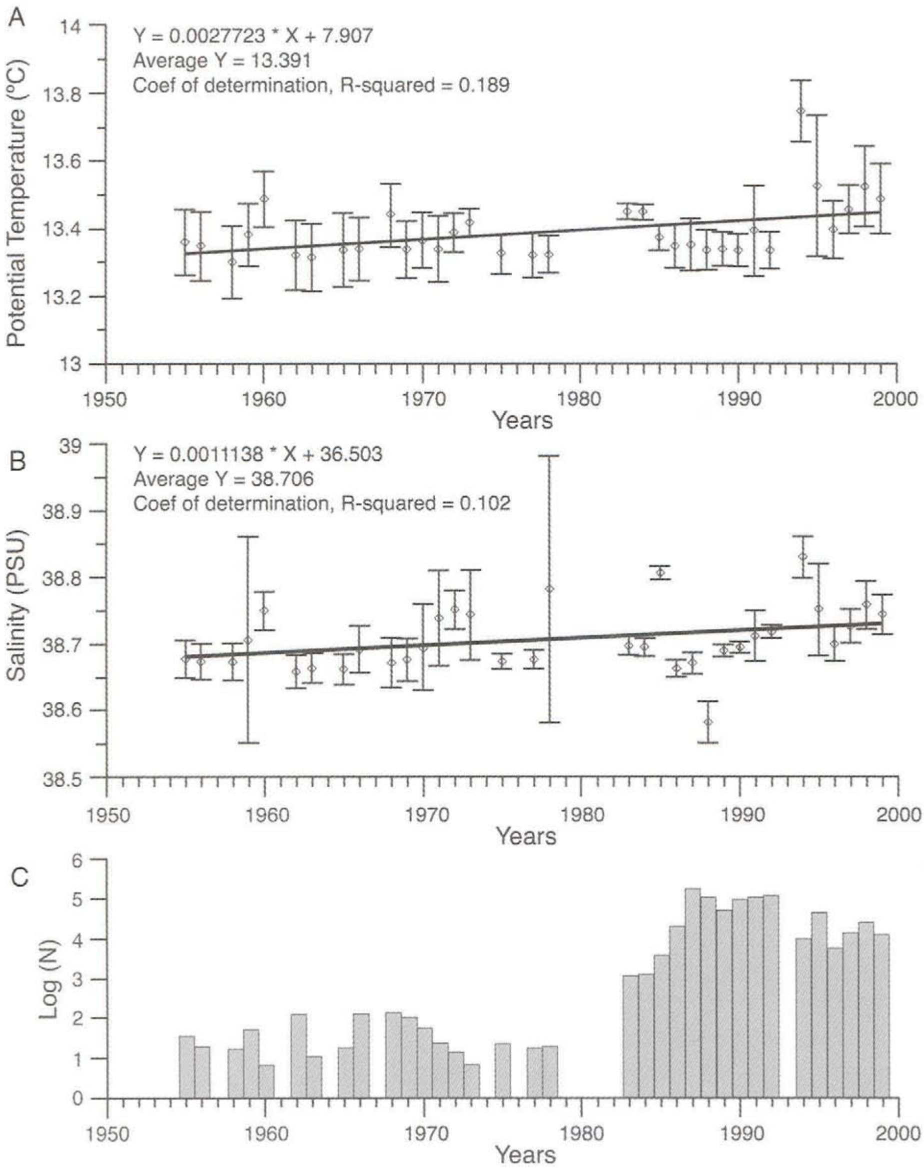


Figure 80: Long-term changes in (A) potential temperature (°C) and (B) salinity of the EMDW (data below 1,200 m) in the Ionian Sea Number of data points (C) in the Ionian deep waters (>1,200 m) obtained by grouping the hydrological profiles annually. The vertical bars denote one standard deviation confidence intervals. The linear regression lines and the R^2 correlation coefficients are indicated. (Figure from Manca et al., 2004; reproduced with permission).

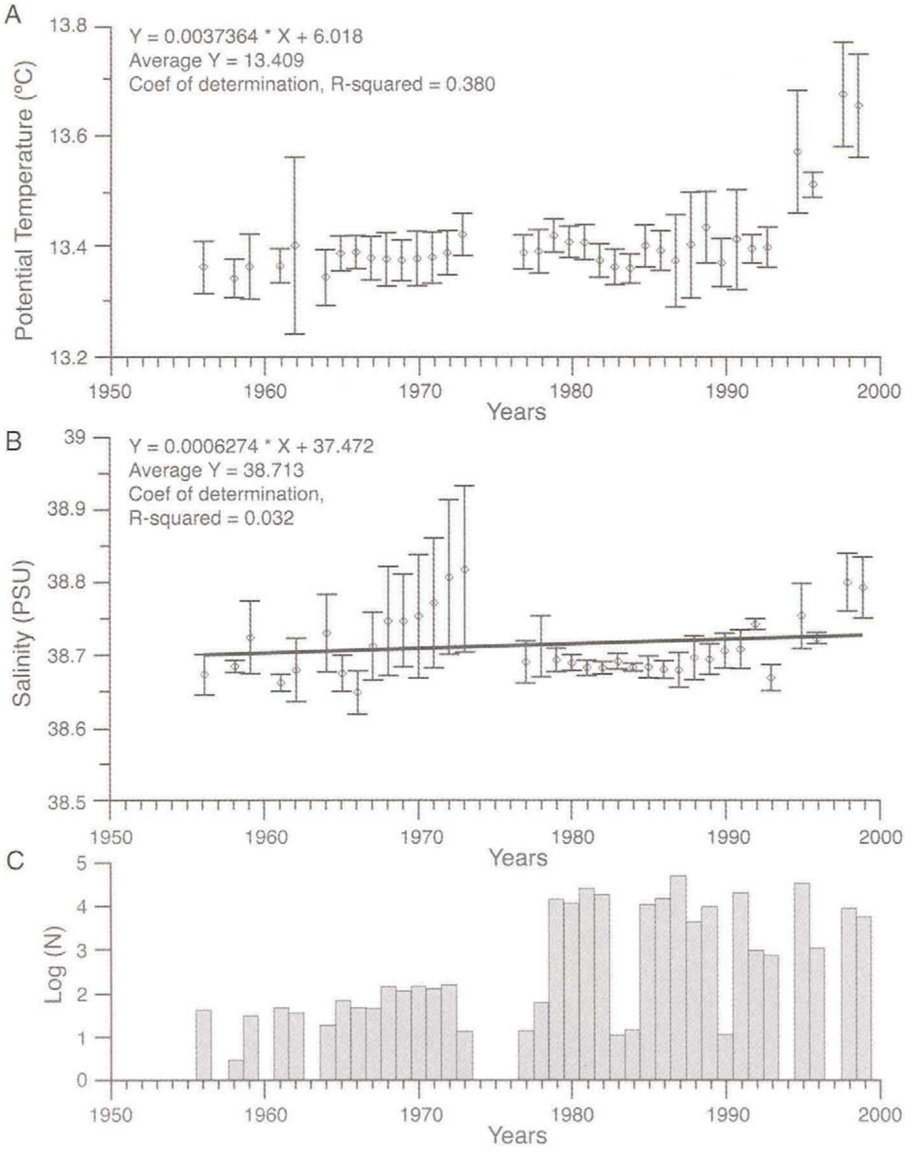


Figure 81: As in Fig. 77, for the EMDW in the Levantine Sea (also from Manca et al., 2004; reproduced with permission).

4.4.2. Rapid Changes: The Eastern Mediterranean Transient

More dramatic than the observed multidecadal trends in T and S is the unexpected change in the location of the East Mediterranean deep water

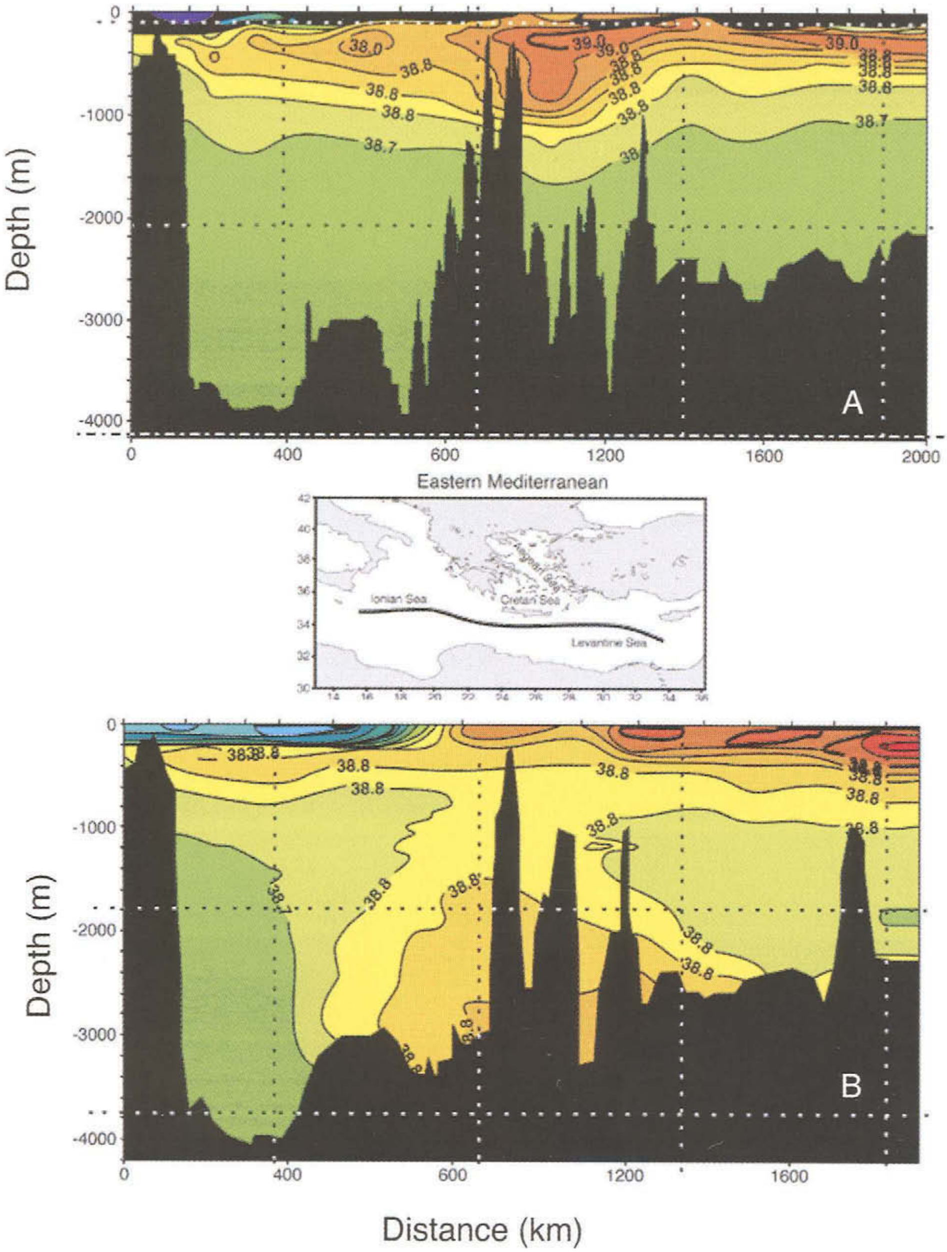


Figure 82: Zonal transect (middle box) of salinity by F/S Meteor south of Crete in (A) 1987 and (B) 1995. Redrawn from Roether et al., 1996.

formation that took the oceanographic community by surprise in the 1990s. The classic view of the oceanic circulation in the Mediterranean Sea had the Adriatic Sea as the major source of the deep waters (EMDW) of the Eastern Mediterranean (Pollack, 1951 and see Section 4.3.1). However, comparison of T and S sections as well as CFC-12 observations south of Crete from two F/S Meteor cruises in 1987 and 1995 (Roether et al., 1996) revealed a dramatic change of the vertical structure of the deep water column of the Eastern Mediterranean (Fig. 82). New, highly saline and rich in oxygen waters filled the bottom layers of the Ionian and Levantine Basins. These water masses were outflowing from the Straits of the Cretan Arc, "pushing" the older EMDW water mass to shallower depths. CFC concentrations clearly revealed the young age of these waters. Roether et al. (1996) estimated an average rate of outflow of dense Aegean waters from the Cretan Straits of about 1 Sv for seven years (1987–1995) although it is not clear from that publication how this number was derived. The new observations shook the physical oceanographic community of the Mediterranean, as they showed that significant changes in the functioning of the thermohaline circulation could occur rapidly. The Roether et al. (1996) publication was preceded by an announcement by Della Vedova et al. (1995) on changing heat transfer between the water column and the sediment, suggesting changes in the near-bottom temperature profile (also, Della Vedova et al., 2003).

The evolution of this change which was termed the Eastern Mediterranean Transient was further clarified through analyses of various oceanographic cruises that took place in the region in the years between 1987 and 1993. The Cretan Sea gradually filled with newly formed, more saline water, with density exceeding 29.2 kg m^{-3}. (Fig. 83, Theocharis et al., 1999b) at a rate reaching 3 Sv in 1991–1992. This water overflowed from the Cretan Straits filling the deep Eastern Mediterranean Basin. Estimates of the outflow based on current meter measurements covering 150 days in 1994 suggested rates close to 0.5 Sv (Tsimplis et al., 1997, 1999). Since then, the signature of the Aegean waters has propagated in the Ionian Sea (Theocharis et al., 2002; Manca et al., 2003). After the mid-nineties, the Cretan Sea returned to the pre-EMT condition of exporting small amounts of dense water that does not reach the bottom of the Ionian and Levantine Basins, but ventilates the depths of 1500–2000 m (Theocharis et al., 2002), while the main contribution of dense water for the Eastern Mediterranean is once again the Adriatic Sea (Klein et al., 2000).

4.4.3. What Caused the EMT?

Major research efforts have been made during the last decade in order to understand the generation, development and demise of the EMT. Many issues

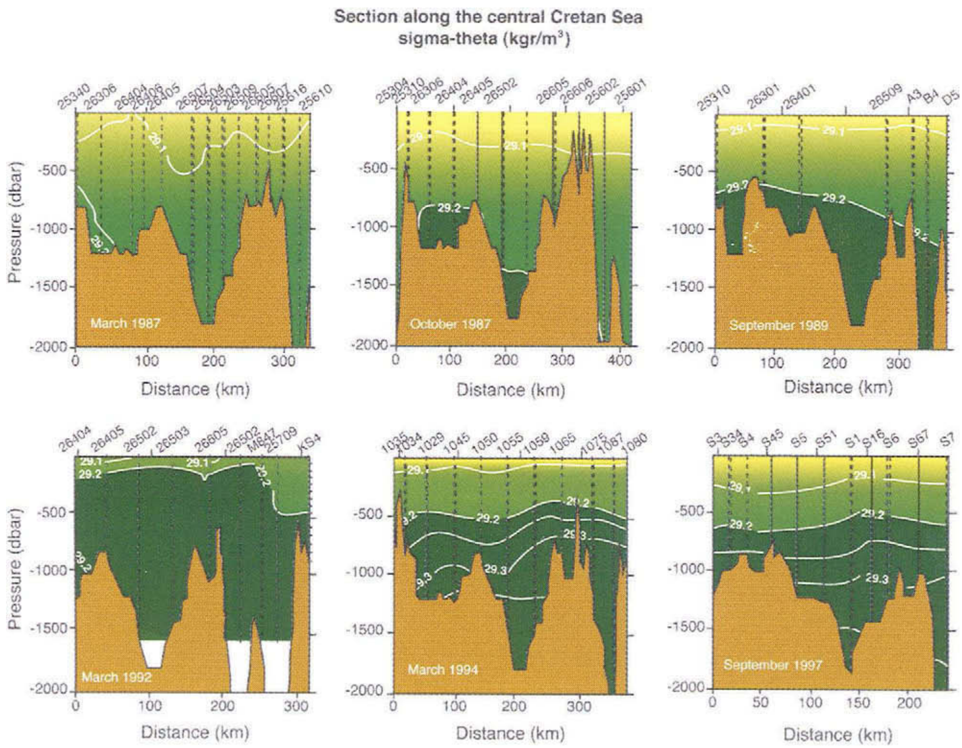


Figure 83: Evolution of the pycnocline in the Cretan Sea from March 1987 to September 1997 (from Theocharis et al., 1999b; reproduced with permission).

have been clarified but also alternative suggestions have been developed. We will try to explain the varying views in respect of the generation and development of the EMT in the context of four questions: When did the EMT start? Where did it start? How much deep water did it produce? and Why did it happen? The answers to these questions are interlinked and determining the location of the EMT also restricts to an extent the start time, the forcing and the quantity produced.

When did the EMT Start?

Several authors state that the deep water formation of 1987 was the beginning of the EMT (Klein, 1999; Lascaratos et al., 1999; Malanotte-Rizzoli et al., 1999; Theocharis et al., 1999b; Zervakis et al., 2000; Nittis et al., 2003). There is also general agreement between these authors as well as researchers looking at the atmospheric forcing (Josey, 2003; Jacobeit, 2005; Tragou et al., 2005), that the major deep water formation incident happened during the very severe winters of 1992 and 1993. However, the 1987 measurements by Roether et al. (1996)

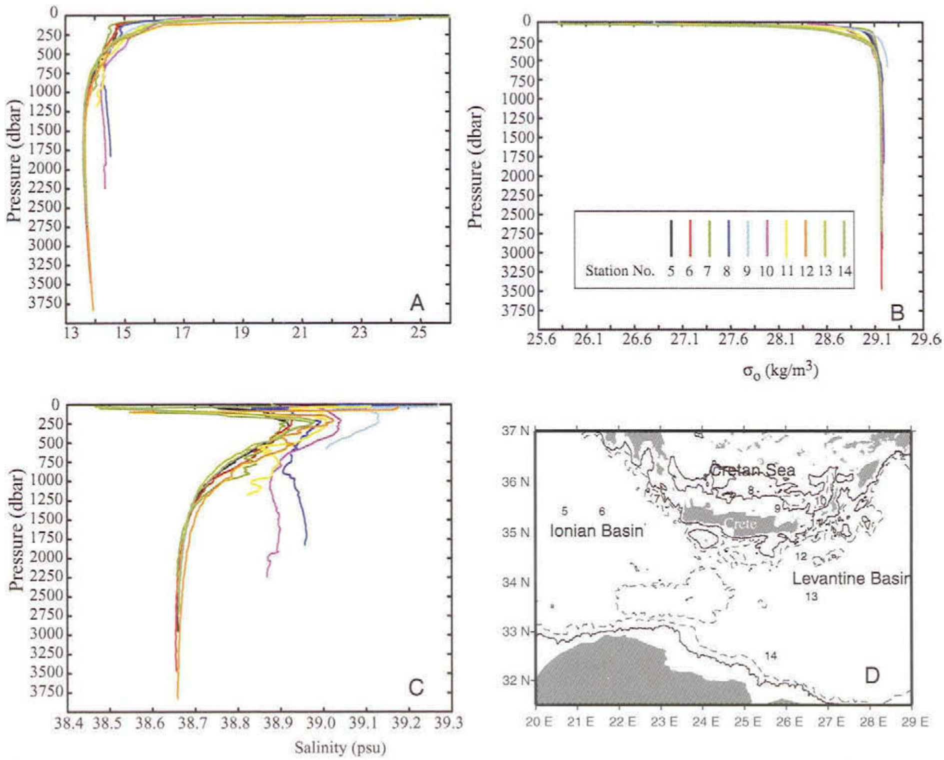
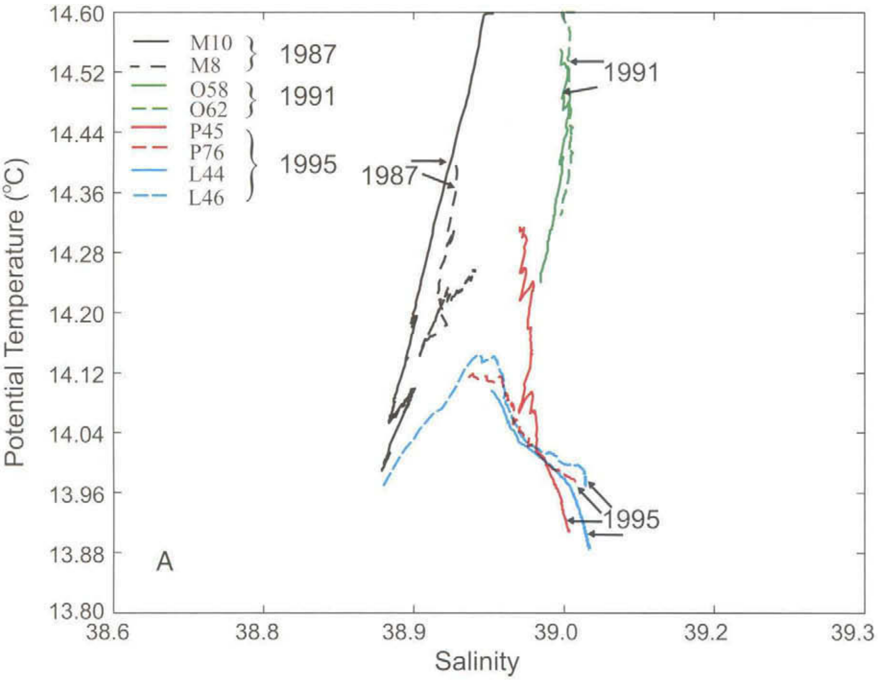


Figure 84: The temperature (A), density (B) and salinity (C) profiles from the 1987 Meteor cruise together with the location of the stations (D). Note the salinity profiles at stations 8 and 9 in the Cretan Basin.

which are generally considered as the background condition before the EMT has started for the deep basin outside the Cretan Sea indicate that within the Cretan Sea dense waters were already forming (Fig. 84).

The development of the water mass properties within the Cretan Basin can be seen in Fig. 83 (from Theocharis et al., 1999b) and Fig. 85 (from Tsimplis et al., 1999) where it is clear that the deep waters within the Cretan Basin became saltier first (between 1987 and 1991) and then (between 1991 and 1995) saltier and cooler. Thus, it appears that, as far as the Cretan Basin is concerned, the deep water formation was not an isolated incident but rather a prolonged one with at least two steps. The two-step scenario has also been suggested by Lascaratos et al. (1999) and Theocharis et al. (1999b). The first step involves increase in salinity and the second one is associated with cooling and further salinity increase. The salinity profile at station 9 (Fig. 84) from the 1987 Meteor cruise indicates that saltier waters were already forming near the continental



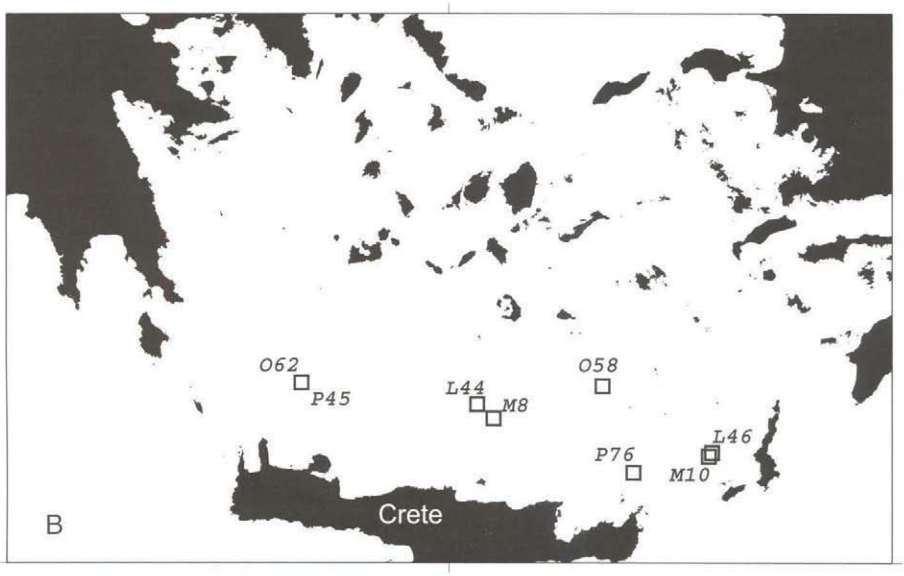


Figure 85: Temperature and salinity plots at selected stations showing the stages of development of the EMT (Tsimplis et al., 1999).

shelf of Crete at that time. These observations suggest a number of possibilities: (1) Deep water formation was not the exception but the norm during, at least the winters of the period 1987–1995 although in some years they may have resulted into smaller volumes of deep water which remained undetected. (2) The 1992–93 deep water formation was linked to the atmospheric forcing identified by Theocharis et al. (1999b) and Josey (2003), but deep water formation had started earlier, by other processes affecting mainly the salinity of the upper waters. These could be linked to river damming (Rohling and Bryden, 1992; Boscolo and Bryden, 2001) or to the continuous reduction of freshwater at the Eastern Mediterranean through the shift of the NAO to an almost permanent high state during the last few decades (see for example, Tsimplis and Josey, 2001) or due to reduction of freshwater in the Northern Aegean because of reduced outflow from the Black Sea (Zervakis et al., 2000). If only the surface salinity is increased then the maximum density will be formed at higher temperature which is what we observe in Fig. 85 in agreement with Tsimplis et al. (1999). Thus we think that long-term processes contributed in the preconditioning of the EMT and the water mass characteristics of the deep water formed.

Where did the Deep Water Formation Take Place?

This is clearly a surprise question for non-experts in Mediterranean Sea oceanography. If the above-made suggestion that the long-term changes in freshwater fluxes contributed to the EMT is correct, then the spatial distribution of such changes must, of course, be distributed to the Eastern Basin through the oceanic circulation. Thus, it is in respect of the events triggering the massive deep water formation that the views in respect of the position of the EMT differ. Clarifying the location is, of course, tied to the identification of the cause and understanding the sensitivity of the basin to local changes in the freshwater and heat forcing.

The first studies of the EMT suggested the Cretan Sea as the sole area of deep water formation (see for example, Roether et al., 1996; Klein et al., 1999; Lascaratos et al., 1999; Malanotte-Rizzoli et al., 1999; Theocharis et al., 1999b) arguably because this was the place where the changes in the deep water characteristics were first observed. Accordingly, most modelling studies have concentrated on exploring the contribution of the atmospheric forcing on the Cretan Sea and its circulation changes. These studies developed the dominant school of thought which suggests that changes in the wind field caused diversification of the MAW to the north of the Ionian Sea (Malanotte-Rizzoli et al., 1999) thus depriving the Levantine from fresher water and led to the generation of the EMT through several circulation changes discussed in Section "Why did the EMT happen?" (Malanotte-Rizzoli et al., 1999). According to this view, the Cretan Sea is the place of deep water formation during the EMT.

The suggested alternative mechanism was a reduction of the Black Sea freshwater into the Northern Aegean (Zervakis et al., 2000). Normally, this water mass being less saline than the underlying water masses acts as a lid that absorbs atmospheric forcing thus not permitting the underlying denser water to be exposed to the highest heat losses directly and thus form deep water. However, Zervakis et al. (2000) observed that the published minima in the Black Sea outflow coincide with the deep water formation periods and are also supported by the available oceanographic observations in the North Aegean. Thus, they suggest that deep water formation started in the North Aegean Basin and then overflowed over the Cyclades Plateau into the Cretan Sea and contributed in the preconditioning for the 1992–93 event. It is worth noting that the deep basins of the North Aegean underwent full replenishment of their waters in March 1987. It is of course possible that both the above mechanisms were triggered in parallel but clearly the relevant importance of each of them is crucial in understanding the system.

How Much Deep Water was Produced by the EMT?

The characterization of the EMT is mainly linked to the outflow of waters over the sills of the Cretan Straits. The initiation of outflow requires production of deep water in quantities capable of filling the Cretan Basin to sill level and the outflowing to the deep basins of the Eastern Mediterranean. As an alternative, one can consider that small amounts of deep waters were formed year after year either until the dense waters of the Cretan Basin reached the sill level and then a much smaller deep water production initiated the outflow. Roether et al. (1996) have estimated the EMT outflow as 1 Sv over the period 1986–1997 but without providing a description of the estimation or any error bars. Theocharis et al. (1999) suggest that the spring and summer outflow must be stronger than the outflow in autumn and winter as deep water formation is a late winter event. Moreover, it is also reasonable to assume that the flow must be reducing as the head of the denser water is reduced. Thus, during the spring following the deep water formation the outflows may have been higher than 1 Sv. However, estimates of the outflow from subsequent current meter measurements suggest lower overall values. A simple hydraulic model estimating the flow by reference to the inclination of isopycnals over the sills (Tsimplis et al., 1999) give values about half those given by Roether et al. (1996). Consistency between the current meter measurements and the estimate of Roether et al. (1996) can be achieved if one accepts that the suggested higher outflows during spring time happened only during the years of deep water formation, namely during 1987, 1992 and 1993 (Theocharis et al., 1999). If 1 Sv is the average over 7 years and we accept that in total the observed values from the various current meter measurement (Tsimplis et al., 1999) indicate an average of 0.5 Sv during the four non-formation years as well as the autumn and winter of the three formation years then the outflows following the formation periods must be in excess of 2.5 Sv in support of the suggested outflow rates of around 3 Sv during the formation period (Theocharis et al., 1999).

However, if formation started in the North Aegean and filled in that basin first before outflowing over the Cyclades plateau to the Cretan Sea (Zervakis et al., 2000) then arguably even higher rates could be suggested.

Why did the EMT Happen?

We have already discussed several of the scenarios suggested as contributing to the generation of the EMT, namely long-term forcing through freshwater reduction from rivers, the NAO pattern or the Black Sea outflow, anomalous heat fluxes during specific winters, specific changes in the circulation of the MAW and the LIW. We now re-examine them and link them to evidence from modelling and observational studies.

Modelling studies have suggested competing or contributing mechanisms and as they cannot all be correct they cannot be conclusive as to how the EMT has been produced. However, they provide valuable insights on the potential contribution of each mechanism separately and the size of the required changes.

Local (over the southern Aegean) anomalous meteorological conditions may have been partly responsible for the shift of dense-water formation activity from the Adriatic to the Aegean Sea (Lascaratos et al., 1999 supported by observations by Theocharis et al., 1999b). Samuel et al. (1999) showed that the period of increased dense-water formation over the Aegean Sea coincided with change in the wind-driven circulation of the Eastern Mediterranean which supplied more LIW to the Aegean Sea (Fig. 86).

Malanotte-Rizzoli et al. (1999) identify significant changes in the oceanic circulation: (1) a gyre developed in the Ionian Sea which deflected the fresher Atlantic Water from its course towards the Levantine and the Aegean Sea resulting in increased salinities in both basins, (2) gyres developed east of the eastern Straits of Crete that blocked the normal progress of the LIW south of Crete and forced the LIW into moving through the eastern Cretan Straits into the Cretan Sea thus increasing the local salinity, (3) this situation which was observed both in 1991 as well as 1995 (LIWEX Group, 2003) also involved Cretan Intermediate Water (CIW) spreading out from the Cretan Sea towards the Sicily Strait. These changes have been linked to changes in the atmospheric forcing (Pinardi et al., 1997).

Wu et al. (2000), in a model study, have shown that a 2°C SST anomaly for seven years over the North Aegean could generate the outflow of large quantities of dense waters through the Cretan Straits. Such changes have been further supported by Zervakis et al. (2000) who identified two major dense-water formation

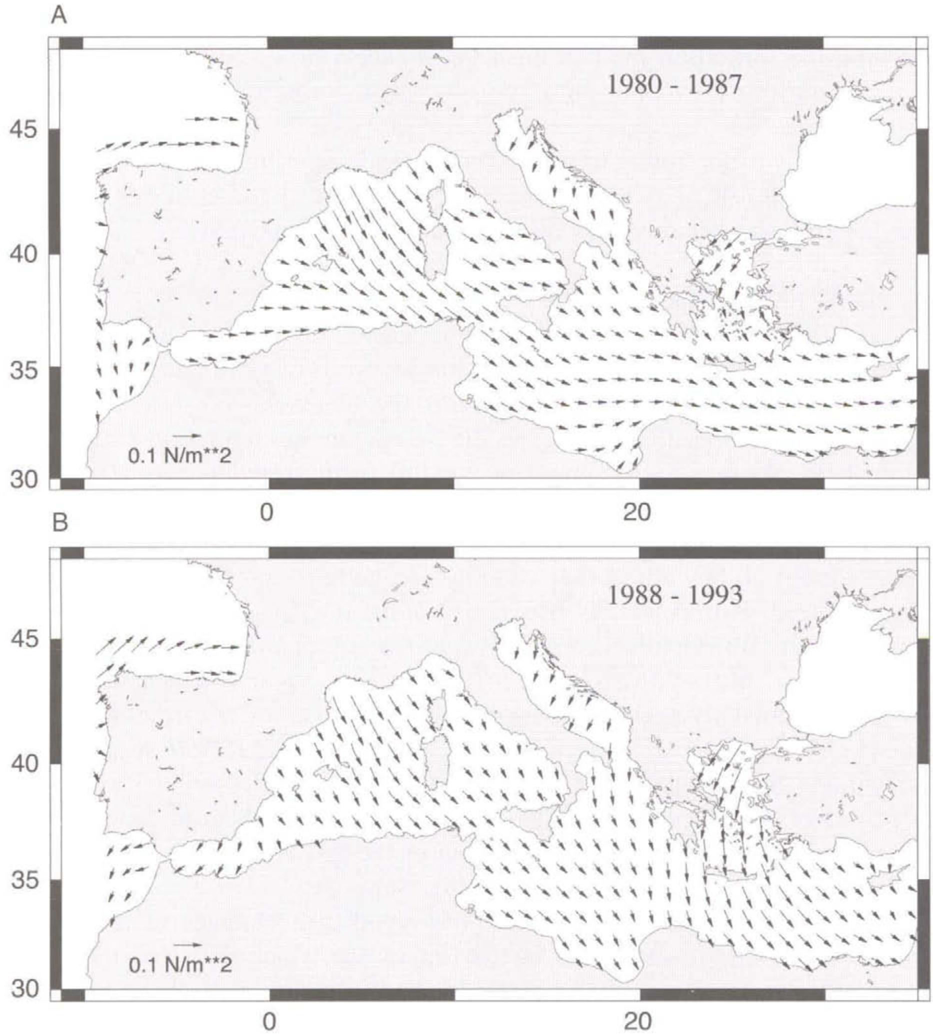


Figure 86: Wind stress from the climatology of the Southampton Oceanography Centre for the periods (A) 1980–1987, and (B) 1988–1993. From Samuel et al., 1999 reproduced with permission.

events in the North Aegean, in 1987 and 1992–93. They proceeded to suggest that the 1987 event was the trigger of the EMT, by accelerating the Aegean circulation "pulling" highly saline waters from the Levantine northwards, thus preconditioning for the second, larger formation event in 1992–93 (Zervakis et al., 2004). In assessing the factors causing the 1987 formation, they suggested that reduced Black Sea outflow could help the erosion of the North Aegean pycnocline. Josey (2003) analysed heat and buoyancy flux data from ECMWF reanalyses and identified anomalously high buoyancy losses from the Aegean in the winters of 1992–93. However, his analysis does not reveal any anomalous air–sea interaction in winter 1987 nor in the rest of the period 1987–1995.

Rupolo et al. (2003), forcing a Mediterranean OGCM with daily satellite SST and ECMWF wind stress for the years 1988–1993, have reproduced the main characteristics of the EMT showing that the phenomenology of this event can be separated in two distinct periods as suggested in Section "When did the EMT start?" During 1988–1991, a wind-induced modification of intermediate layer circulation caused an increase in salinity in the Aegean Sea and a contemporary decrease in the inflowing intermediate water in the Adriatic. These changes in the intermediate circulation make the Aegean Sea the favourite site for the production of dense water and the EMT fully develops during the cold winters 1992 and 1993 when the dense water of Aegean origin fill the deep layers of the Ionian Sea, substituting the "old" ADW. Recently, data provided by ex-USSR cruises in the Aegean suggest that the pre-1990 deep-water formations took place north of the Cyclades (Gertman et al., 2005) and support the Malanotte-Rizzoli et al. (1999) scenario of salinification of the Aegean through blocking of the MAW entering the Levantine Sea by the development of a gyre in the north Ionian Sea.

Boscolo and Bryden (2001) suggest that the damming of the Nile and the ex-USSR rivers flowing into the Black Sea lead to an effective increase of the freshwater deficit over the Eastern Mediterranean. This process could erode the stratification in the Cretan Sea and produce dense water thereby initiating the EMT. However, river outflow measurements do not support long-term reduction in the river outflow of the Black Sea (Stanev and Peneva, 2002). On the contrary, observed reduction of the evaporation is likely to have led to increased freshwater outflow in the Aegean Sea (Tsimplis et al., 2004). Skliris and Lascaratos (2004) based on a model analysis, find that a significant part of the observed salinity changes could be due to the effect of damming of the Nile thus implying significant preconditioning due to this anthropogenic cause. However, they also note that the effect of damming is probably smaller than the effect of long-term changes in the E-P balance (Skliris and Lascaratos, 2004). Tsimplis and Josey (2001) suggest that the NAO-induced changes in oceanic E-P as well as river outflow is larger than the suggested freshwater input reduction cause by the damming of the Nile.

The relative importance of long-term slow changes like the damming of rivers and the NAO index, against seasonal processes like abnormally cold winters, and short timescale processes such as extremely cold days associated with relatively rare synoptic features remains an open issue in the debate of the EMT. Similarly, it is unclear whether localized changes like a sudden reduction of the Black Sea outflow, or deprivation of the Eastern Mediterranean from freshwater through the damming of the Nile are more important than the basin overall water and salt budget. Whether the forcing was caused by blocking of the Atlantic inflow or was due to local (in the southern Aegean and Levantine) processes or to processes in the North Aegean needs also to be clarified. Thus, the studies of the EMT challenge our knowledge about both the temporal and spatial scales involved and our ability to separate and quantify the contribution of the various forcing mechanisms.

4.5. The Impact of Large-Scale Atmospheric Variability on the Mediterranean Sea

Large-scale patterns of atmospheric variability, for example the North Atlantic Oscillation (NAO), that have their primary centres of action over neighbouring ocean basins have the potential to exert an influence on the Mediterranean Sea. It is well known that the NAO has a significant impact on Mediterranean climate, in particular the amount of precipitation (e.g. Hurrell, 1995) the river runoff (Struglia et al., 2004) and the sea level variability (Tsimplis and Josey, 2001). Moreover, a sudden change observed in the LIW characteristics over the winters of 1981 and 1982 (Brankart and Pinardi, 2001) was attributed to surface heat budget changes over the basin. Rixen et al. (2005) suggest an NAO influence on the Western Mediterranean Deep Water similar to that effected by the NAO in the North Atlantic. They also suggest that the anticorrelations found between the Mediterranean SSTs and the NAO as well as the anticorrelation found by Tsimplis and Rixen (2002) between the upper water temperatures in the Adriatic and the Aegean Seas and the NAO has probably played an important role in establishing the Eastern Mediterranean Transient (Demirov and Pinardi, 2002).

However, the extent to which large-scale patterns play a role in the variability of the ocean circulation, in particular the major EMT event discussed in Section 4.4 is not well established. Some progress has been made and the link between surface forcing and the EMT, together with its possible relationship to large-scale climate patterns has been discussed earlier. Tsimplis and Josey (2001) have suggested a link between the long-term changes in the deep water of the Mediterranean, the observed sea level changes and the development of the EMT with the consistently high values of the NAO index during that period. Although they were justified in their assessment about the sea level changes and their assertion about the NAO effect in respect of the deep water changes is also shared by other researchers (see for example, Millot, 1999; Send, 1999) their suggestion about the link between the NAO and the EMT appears less strong. The latter suggestion was based on their observation that between 1988–93, the freshwater changes due to changes in the E-P which peaked over the

Aegean and changes in the river outflow were sufficient to cause, at least partly the observed freshwater deficit that has been suggested as a cause of the EMT. However, subsequent work by Josey (2003) suggests that the E-P contribution was probably not significant while Tsimplis et al. (2004) estimate, on the basis of an analysis of sea level variations in the Black Sea an increasing outflow of Black Sea water due to local reduction in evaporation over the Black Sea. Thus, it presently appears that any contribution from the NAO is likely to be affecting the long-term preconditioning of the waters only (as suggested by Tsimplis and Rixen, 2002 and Rixen et al., 2005) but it did not trigger the deep water formation incidents of 1991–92 and 1992–93. In contrast, Josey (2003) suggests that the very cold winters during which the EMT took place are linked to anomalously high pressure over Western Europe and the North-East Atlantic thus implying another major mode of atmospheric variability which is potentially related the East Atlantic Pattern (EAP). This suggestion is supported also by Dunkeloh and Jacobeit (2003) and Jacobeit (2005) who identified a significnat change in the second canonical correlation pattern of geopotential height and precipitation after 1987 (Fig. 74 from Dunkeloh and Jacobeit, 2003). However, it is clear that the EMT has started before 1991 thus other factors must have contributed to its initiation as there were no significant thermal or haline forcing anomalies prior to this time (Josey, 2003). The extent to which these factors are linked with large-scale atmospheric patterns or were just coincidental is not presently resolved.

Is the EMT a unique phenomenon or has it been happening in the past without us noticing? Some suggestions exist for an earlier EMT in the very noisy hydrographic data of the Ionian Sea. Moreover, extreme winter heat loss also occurred in the mid-1970s and this could have been expected to produce conditions conducive to deep water formation in the Aegean Sea (Josey, 2003). However, the significance of this earlier extreme forcing is complicated by long-term river runoff trends (Rohling and Bryden, 1992) which may have led to stronger stratification in the Aegean Sea and greater resistance to convection at this time. Thus it presently remains unclear whether an event similar to the EMT occurred during the mid-1970s.

4.6. Sea Level Changes in the Mediterranean Sea

The Mediterranean sea level depends on the pressure gradients across the Strait of Gibraltar, the prevailing hydraulic conditions at the Strait (Ross et al., 2000), the steric variations and changes in the water budget driven by the regional atmospheric forcing (Tsimplis and Josey, 2001) and the circulation within the basin. A few long sea level records spanning to the beginning of the 1900s exist in the Mediterranean Sea and these are located at the northern coasts of the Western Mediterranean (Marseille and Genoa) and at the northern coasts of the Adriatic Sea (Trieste and Venice) (Tsimplis and Spencer, 1997; Tsimplis and Baker, 2000). The sea level trends for the three longer stations are in the range 1.1–1.3 mm/year, that is, slightly lower than the estimated global value for sea level rise which is in the range of 1–2 mm/year (Church et al., 2001). A fourth long-term tide gauge record in Venice has particularities due to variations in local subsidence and therefore it is not usable in the context of sea level trends during the period when ground water extraction cause local subsidence (Woodworth, 2003). New data for Trieste extending the record for this tide gauge fifteen years back to 1875 leave the above estimates unaffected (Fabio Raicich, personal communication).

The Mediterranean sea level variability can be separated into three periods. Until the 1960s, sea level in the Mediterranean Sea had trends equivalent to those at the open ocean stations (Tsimplis and Baker, 2000). During the second period, between 1960 and the beginning of the 1990s sea level in the Mediterranean Sea was either not changing or decreasing (Orlic and Pasaric, 2000; Tsimplis and Baker, 2000) mainly due to atmospheric pressure changes during the winter period (Tsimplis and Josey, 2002; Woolf et al., 2003) as well as temperature (T) reduction and salinity (S) changes linked to the NAO (Tsimplis and Rixen, 2002). It appears that these T and S changes are restricted to the northern part of the basin, namely the Aegean and Adriatic Sea (Painter and Tsimplis, 2003) while there are basin-wide east–west gradients on atmospheric pressure as well as E-P (Tsimplis and Josey, 2001).

The third period of interest between 1993 and 2002 is based on analyses of the TOPEX/POSEIDON dataset (Cazenave et al., 2001; Fenoglio-Marc, 2002) which reveal a picture much more complicated than that of a coherently varying basin. During this period fast sea level rise was observed at the Eastern Mediterranean Sea (Cazenave et al., 2001; Fenoglio-Marc, 2002) and was linked with changes in observed sea surface temperature (Cazenave et al., 2001). Recently, on the basis of altimetric data, Vigo et al. (2005) confirmed an abrupt reduction of sea level rise rates (Fenoglio-Marc, 2002) as well as negative trends in parts of the Eastern Mediterranean Sea after 1999. These changes appear consistent with sea surface temperature changes and it is suggested by Vigo et al. (2005) as being a consequence of the restoration of the Adriatic Sea as the main source of deep water in the Eastern Basin. The observed sea level values and their temperature-related forcing has been confirmed by the use of climatological data of oceanic temperatures (Tsimplis and Rixen, 2002) for part of the 1990s.

During the same period of time a reduction in the sea level gradient across the Strait of Gibraltar has been observed and the change has been suggested as caused by varied hydraulic conditions in the Strait (Ross et al., 2000) or by changes in the density difference between the Mediterranean and the Atlantic (Brandt et al., 2004). However, a reverse jump in the sea level slope across the Strait occurs in 1999 (Tsimplis et al., 2005b) in the slope and this is accompanied with the sea level rise reduction or reversal observed in the Eastern Mediterranean Sea (Fenoglio-Marc, 2002; Vigo et al., 2005). Thus it is possible that there is a link between the Eastern Mediterranean sea levels and the hydraulic jumps in the Strait of Gibraltar, possibly through the salinity changes in the LIW (Tsimplis et al., 2005b).

Between 1958 and 2001, the tide-gauge records indicate sea level trends in the range of 0.8–0.4 mm/year $\pm$ 0.4 mm/year (Tsimplis et al., 2005a). During the same period, the direct meteorological forcing caused sea level reduction of -0.4 to -0.6 mm/year linked with the North Atlantic Oscillation and particularly increasing atmospheric pressure. After the removal of the meteorological influence from the sea level records the resulting trends were found to be $\sim 0.3 \pm 0.4$ mm/year at the Western Mediterranean and $\sim 1.3 \pm 0.4$ mm/year at the Eastern Mediterranean which is strongly affected by rapid sea level rise in the period 1993–2001 with rates of 5–10 mm/year probably related to the Eastern Mediterranean Transient (EMT) (Tsimplis et al., 2005a).

In conclusion, during the last 40–50 years sea level trends within the Mediterranean Basin differ significantly from those of the nearby Atlantic Ocean (Tsimplis and Baker, 2000; Woolf et al., 2003). It is yet unclear for how long can the Mediterranean Sea sustain sea level behaviour different from the open ocean. To the extent that the differences are caused either by large-scale meteorological patterns like the NAO or basin or sub-basin steric processes which do not affect the pressure gradient across the Strait of Gibraltar, substantial differences may be permitted to remain. However, if a mechanism of mass addition to the open ocean by melting ice is assumed as the primary cause of sea level rise as suggested by Miller and Douglas (2004) and provided that such a mechanism is enhanced with time (Church et al., 2001) it is unlikely that the Mediterranean Sea will be able to sustain its distinctive behaviour for more than 20–30 years.

4.6.1. Extreme Sea Levels

Changes in extreme sea levels are arguably more important than mean sea level rise as they pose significantly higher risks to coastal regions. Within the Mediterranean Sea very few studies on extreme sea levels have been conducted and even fewer are concerned with changes in extremes. Moreover, the published studies are not basin-wide but rather regional and limited in scope. Lionello (2005) has analysed the trends of extremes storm surges on the basis of the records in Venice and found no significant trend since 1940 apart from that produced by the combination of sea level rise and local ground subsidence. Raicich (2003) has investigated sea level extremes in Trieste for the periods 1939–2001 and found a decreasing trend in strong positive surges in spite of increases in southerly winds due to increased atmospheric pressure (Raicich, 2003; see also Pirazzoli and Tomasin, 2002, 2003; Trigo and Davies, 2002). Tsimplis and Blackman (1997) have documented the sea level extremes for the Aegean Sea for a period of eight years (1982–1989) but no information on trends of extremes could have been derived with these short time-series. However, Tsimplis and Blackman (1997) suggest that the observed extremes are in most cases common in the whole of the Aegean Basin and are consistent with a linear addition of the extreme pressure and wind effects thus implying that knowledge of these fields would suffice for estimating changes in the sea level extremes in the Aegean Sea.

A global study (Woodworth and Blackman, 2004) suggests that extreme sea levels have been increasing during the last decades at the European coasts and that these changes are consistent with the impacts of the North Atlantic Oscillation. This study had been heavily influenced by the northern European stations and thus its results are not representative of extreme sea levels within the Mediterranean Sea. The mean sea level changes described above would cause extreme sea levels to decline between the 1960s and the 1993 and to increase afterwards. As the NAO remains the major atmospheric forcing pattern during winters until 2001 (Tsimplis et al., 2005a) it is unlikely that significant changes in the maximum surges would have occurred and the changes in the mean sea level which was NAO controlled at the Eastern Mediterranean until 1993 and EMT controlled after 1993 are likely to be the dominant change in the sea level extremes.

The problem of extremes in the Mediterranean Sea is complex. In this book, it is also discussed in Chapter 6, in relation with the cyclones in the Mediterranean region. It is worth noting that floods have been observed with non-exceptional atmospheric conditions in the east Adriatic when resonance between travelling air pressure disturbance with a coastal wave (Vilibic et al., 2004). Moreover in the Adriatic in addition to the contribution of storm surge, seiches and tides, low-frequency oscillations ($0.01 < f < 0.1$ cpd) predominantly induced by planetary atmospheric waves are also important in producing extreme sea levels (Pasaric and Orlic, 2001). Thus, changes in each of the forcing parameters may affect the distribution of sea level extremes. Recently a correlation between sun spot activity and sea level extremes in Venice has been reported (Lionello, 2005) although a plausible physical mechanism for such a link has not yet been suggested. However further work has apparently shown similar results in other European stations during the second half of the last century (P. Pirazzoli, personal communication).

4.7. Changes in the Wind-wave Field

Studies on the changes in the wind-wave fields are hindered by the lack of data. In particular, wave buoy observations are only available in limited locations and only since the early 1990s. Non-directional satellite observations of significant wave height are also available for 1993–2002 as two measurements per day above the Mediterranean Sea. Thus the major source of wave-field studies is model simulations based on the atmospheric forcing derived from reanalyses of atmospheric data. However, such simulations because of their rough resolution are affected by systematic surface wind under-evaluation. This problem affects both the analysis of monthly average SWH (Significant Wave Height) fields and extremes SWH values, which are presented in Section 6.3.4.

The variability of the monthly average SWH (Significant Wave Height) field in the Mediterranean Sea, in the period 1958–2001 has been analysed (Lionello and Sanna, 2005) using the data provided by simulations carried out using the WAM model (WAve Model cycle 4) forced by the wind fields of the ERA-40 (ECMWF Re-Analysis). Comparison with buoy observations, satellite data, and simulations forced by higher resolution wind fields indicate that, apart from the underestimation of the observed SWH, space and time variability of the wave fields are correctly simulated by the models.

The SWH field shows large inter-annual and inter-decadal variability and a statistically significant decreasing trend in mean winter values of (−0.2 cm/year) mainly caused by a weakening of the Mistral wind regime during winter (Fig. 87). When the mean annual SWH values are considered a much smaller trend of −0.08 cm/year is found.

The correlation coefficient between the winter average SWH and the winter NAO index is −0.47 and −0.73 for the seasonal and 5-year low-pass filtered time series, respectively (Fig. 88). In spite of the significant correlation, the link between Mediterranean SWH variability patterns and NAO is not that strong. The SLP composite, based on winter months of the average SWH shows a pattern different from the NAO (Fig. 89, upper panel). This composite has been obtained subtracting the average of the set of fields when the average winter SWH was in the 10% lowermost range to that when it was in the 10% uppermost range. The composite is not similar to the NAO dipole. It presents a tri-pole with a principal minimum located over central Europe, a minor one in the Atlantic and a large maximum over North Atlantic, extending to Russia. On the Mediterranean

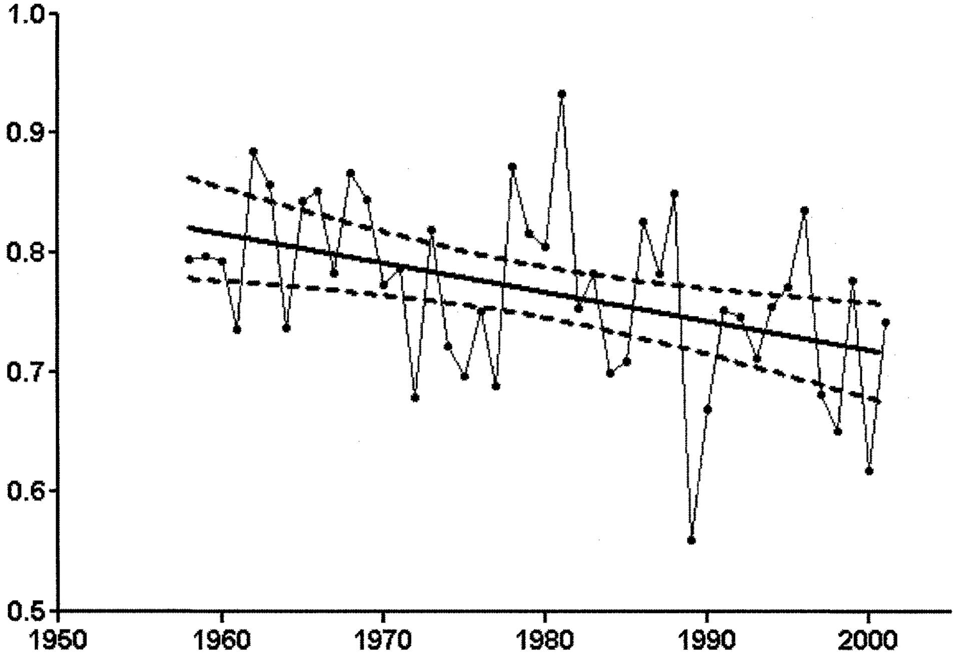


Figure 87: Time series of the average winter SWH (values in meters), linear regression (straight line) and 95% confidence interval (dashed line) (from Lionello and Sanna, 2005).

region, it determines a strong atmospheric circulation, following the shape of the basin, with a prevailing south-eastward direction in the western basin and north-eastward in the eastern one.

During summer, the wave field variability appears weakly correlated with the Indian Monsoon index (Figs. 88, 89, lower panels) with correlation coefficients 0.34 for the summer values and 0.5, which is not significant for the 5-year low-pass filtered time series. This correlation reflects the moderate influence of the Monsoon on the meridional Mediterranean circulation.

Thus the SLP patterns associated with the SWH inter-annual variability reveal structures different from the NAO and Monsoon circulation although these indices are significantly correlated with the winter and summer SWH, respectively. This is because the wave-field variability is conditioned by regional storminess in combination with the effect of fetch. The latter is likely to be the most important. Thus, although the role of the large-scale patterns (mainly NAO) influences the average SWH field, their effect is strongly modulated by mesoscale factors and geographical land–sea distribution. The fetch acts as a filter, selecting surface atmospheric circulation components where regional characteristics

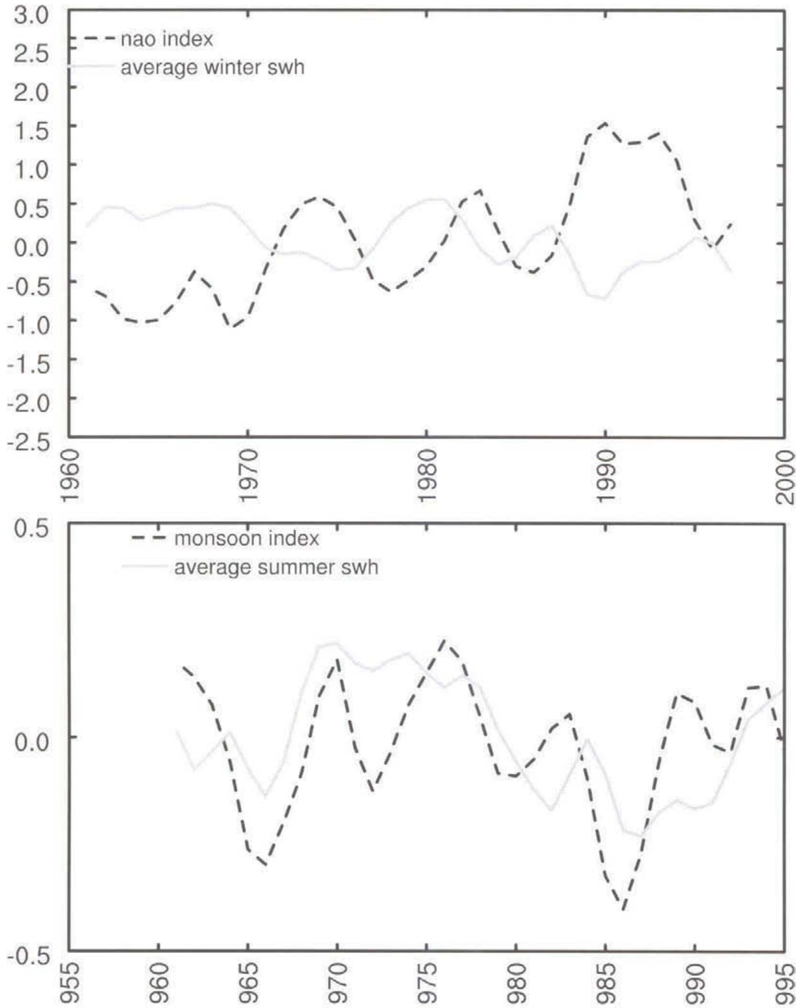


Figure 88: Time series of winter NAO index (dashed black line) and winter SWH index (light grey continuous line) (top). Time series of Indian Monsoon index (dashed black line) and winter SWH index (light grey continuous line) (bottom). A 5-year low-pass filter has been applied to all time series (from Lionello and Sanna, 2005).

conform to the shape of the basin and, therefore, are more effective in producing waves. This modulation implies that variability regimes reflect regional features, directly responsible for the wave generation, more than large-scale patterns, though a link to NAO and Indian monsoon is certainly present.

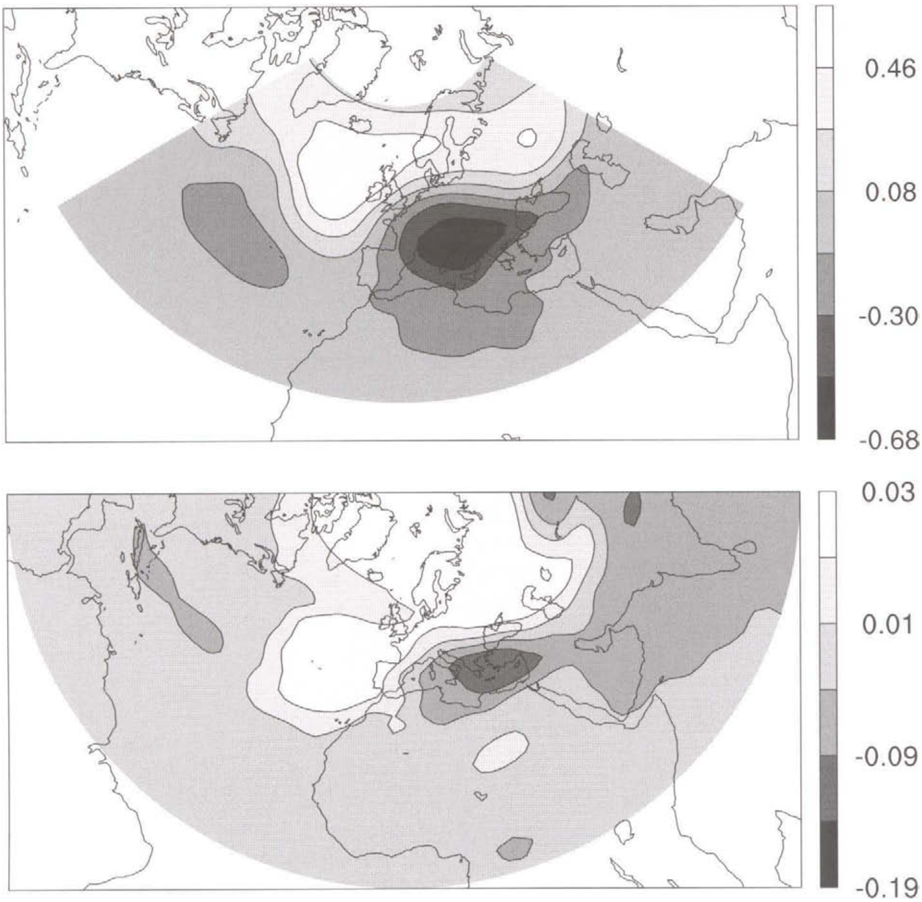


Figure 89: SLP composites based on monthly average SWH deviation from annual cycle for winter (top) and summer (bottom). Contour levels according to the level bars; values in hPa. These composites are obtained subtracting the average of the set of fields when the average SWH was in the 10% lowermost range to that when it was in the 10% uppermost range. Contour levels according to the level bars; values in hPa (from Lionello and Sanna, 2005).

4.8. Outlook and Future Challenges

During the last decade the Mediterranean Sea has been an exceptional place to do oceanographic and climatic research. The slow changes in the water mass characteristics of the deep waters of the Mediterranean Sea and their potential links to atmospheric forcing and/or to damming of the rivers together with the rapid changes in the deep water formation locations as evidenced by the EMT and the uncertainty that exists both on the exact forcing mechanisms as well as the possibility of being a regular pattern which we managed to observe for the first time have raised issues about the coupling between the rapid and the longer term changes, their respective significance and the triggering mechanisms. Moreover, the exceptional character of observed sea level changes which appear not linked with the global sea level situation raise the question whether the Mediterranean Basin future sea level scenarios cannot be based on the global ones as these do not include the relevant forcing mechanisms.

Have these changes affected the Mediterranean society? Estimating the cost of economic or social activities related to the use of the sea by the Mediterranean countries is not easy. Rough estimates of the value of the economic activities by the first author suggest that the total of the sea-related economic activities does not exceed 5% of the GDP for each country.

In addition to the economic implications for society, ecosystem changes may become important. The ecosystem of the Mediterranean and the Black Seas has been characterized as "sensitive" and the future not "rosy" (Turley, 1999). The major influence appears to be linked with eutrophication caused by increased agricultural phosphates and the damming of rivers (Turley, 1999) rather than to climatic changes. Dust deposition which carries Fe and is very important for phytoplankton growth is, according to Turley (1999), controlled by the NAO. Consequently, the varying dust input is likely to affect climate change as well as the availability of nutrients in the basin although the results are presently unpredictable (Turley, 1999). The multiscale variability of the Mediterranean Sea has serious implications for policy making and environmental management (Zavatarelli, 1999). It is of the highest priority for the scientific community to assess the impact the observed oceanographic changes have on the Mediterranean society and prioritize in accordance with these impacts. Preliminary assessments of some specific processes like the effect of climate change on estuaries have been published for particular countries (Paskoff, 2004) and national impact assessments are presently underway is several countries (Nadia Pinardi and Enrique Alvarez Fanjul, personal communications). It is therefore necessary to develop a Mediterranean Impact Assessment study exploring the vulnerability of the Mediterranean Sea natural and socio-economic system. This would require interaction with social scientists and engineers as well as with governing bodies which will need to assess the efficiency of their decision-making mechanisms. Whether this is a feasible process for all Mediterranean countries is unfortunately doubtful due to the regional geopolitical complexities. Thus it is important that the developed Mediterranean countries support the developing ones in establishing monitoring systems at their coasts as well as performing their national impact assessments.

The developments in satellite oceanography has led to the realization that smaller scale processes are present and need to be accounted for in order to understand changes in the system. Thus present efforts of long-term monitoring should be extended and improved (CIESM, 2002).

The identification of the main modes of temporal and spatial variability of circulation in the Mediterranean Sea and their response to different modes of atmospheric variability are not yet fully resolved. Within this context the analysis the relationship between changes in the Mediterranean circulation, deep water formation, sea level and regional climate variability on the one hand and local synoptic scale processes on the other hand should be promoted.

Exploratory assessments of future changes in Mediterranean circulation, water mass characteristics and sea level under scenarios of a warming atmosphere would be useful tools both scientifically and for policy-making reasons.

Similarly for the Black Sea there are still many issues to be addressed. The research is to be directed towards interdisciplinary issues. Coupled models (atmosphere–ocean, or physical–biogeochemical models) have to address the various feedbacks between the components of the climate and biogeochemical systems. For the numerical modellers the challenge should be to unify the adjacent basins (Mediterranean, Marmara, Black and Azov Seas) addressing the climatic controls of straits.

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