The 1987 Aegean dense water formation: A streamtube investigation by comparing theoretical model results, satellite, field, and numerical data with contourite distribution

M. Bellacicco a,b, C. Anagnostou c, F. Falcini a, E. Rinaldi a, K. Tripsanas d, E. Salusti a,*

a Istituto di Scienze dell’Atmosfera e del Clima (ISAC)-CNR, Via Fosso del Cavaliere 100, Roma, Italy
b Università degli Studi di Napoli “Parthenope”, Via Ammiraglio Acton, 38, Napoli, Italy
c Hellenic Centre for Marine Research, Institute of Oceanography, 46.7 km Athens-Sounio Av., 19013 Anavyssos, Greece
d Geology Home Team, Shell U.K. Limited, Aberdeen, AB 12 3FY, United Kingdom

ABSTRACT

We here discuss a detailed investigation of the dense water formation, evolution and spreading in the Aegean Sea during the year 1987, immediately prior to the onset of the Eastern Mediterranean Transient (EMT). We use hydrological data collected during the LIA cruise; satellite images for SST (Sea Surface Temperature), and PROTHEUS data (a coupled ocean–atmosphere numeric model) along with theoretical streamtube models. These hydrological analyses are related to late Quaternary sedimentary drifts in the Cyclades Plateau and in the Myrtos Basin. Our analysis shows that streamtube dynamics provide a novel model of dense water evolution and spreading in the Aegean Basin. Applying this model to dense water masses observed in winter and spring 1987 near Samothrace and over the Limnos-Lesbos Plateau, results in a geostrophic flow of this dense, cold water towards the Limnos-Sporades Channel, in the North Aegean Sea. There it mixes with dense water from the Limnos-Lesbos Plateau and finally both move geostrophically towards the Cyclades Plateau. These results indicate that most of the dense water observed near the Cyclades, formed initially about 3 months earlier at Samothrace and Limnos shelves. During its long pathway it partially mixed with adjacent water masses. Although our analysis concerns only one year of dense water analyses, these results are thought to reflect a more general and recurrent phenomenon in the Aegean basin. Indeed, high-resolution (Airgun 10 in.) seismic-reflection data from the Cyclades Plateau reveal the presence of late Quaternary sediment drifts. These observations are concordant with results from our theoretical model. This suggests a direct link between such a dense-water cascading and contourite dynamics. The continuation of sediment drifts into the deep basin floor (≈ 900 m deep) of the Myrtos Basin, moreover, indicates a cascading character of such bottom currents at the flanks of the basin, a feature that set further investigations.

© 2016 Elsevier B.V. All rights reserved.

1. Introduction

The idea that the Aegean Sea could be the source of the Eastern Mediterranean deep waters was first proposed by Nielsen (1912), but successive observations demonstrated the main role of the Adriatic Sea as source of such bottom waters (Malanotte-Rizzoli and Hecht, 1988, and references therein). Field observations from the Aegean Sea around the year 1990 revived the Nielsen hypothesis by demonstrating that the Cretan dense water (CDW) was observed on the Northern Aegean coasts and on the shallow shelves of the Cyclades islands, and then accumulated in the Cretan Sea (Zervakis et al., 2000; Gertman et al., 2006). CDW, later on, filled the bottom layers of both Mediterranean Sea basins (Roether et al., 1996; Klein et al., 1999) although, overflowing the Sicily Strait, this deep water results to be heavily modified (Ben Ismail et al., 2014).

During 1987, before the EMT event, an unexpected dense water was observed in the Aegean Sea. Its density (σθ ≈ 29.25–29.35) was indeed higher than the historical maximum value of 29.2 (Georgopoulos et al., 1992; Theocharis and Georgopoulos, 1993, two similar articles called G2T3 in the following).

Whereas the period 1992–1993 is generally considered as the real onset of a full EMT (Roether et al., 1996), the cold winter of the year 1987 is of some interest since modern field data are available and also because of a particularly severe 3–13 March storms (Lagouvardos et al., 1998) due to cold fronts passing over the Aegean Sea.

About the EMT event, during September 1987 the first Meteor-POEM cruise did not find any presence of the CDW outflow into the deep Eastern Mediterranean, while Gertman et al. (2006) evidenced a CDW outflow from the Island of Kassos in the Cretan Arc (Fig. 1) during...

We here focus on path and formation of dense water currents during the 1987 cold winter and spring, in order to infer recursive phenomena of geological interest. We pair field data (i.e., hydrographic data from G2T3) with theoretical models (i.e., the streamtube model by Smith, 1975; Killworth, 1977; Rydberg, 1980, and other viscous models by Shaw and Csanady, 1983; Shapiro and Hill, 1997; Killworth, 2001), numerical outputs (from the PROTHEUS numerical model; Artale et al., 2010), and satellite measurements (AVHRR time series; Josey, 2003; Marullo et al., 1999a, 1999b) to infer dense water formation sites. Such an approach aims to determine the possible pathway and spreading of the resulting bottom currents from the Northern Aegean Trough and the Limnos-Lesbos Plateau till the Cyclades Plateau, also in relations with contourite deposits in this plateau. All this is complemented by the analysis of a large set of high-resolution seismic-reflection data from the Cyclades Plateau and Myrtoon Basin in order to map late Quaternary sedimentary structures. The combination of marine geological data with oceanographic analyses provides relation between bottom-current activity and contourite drifts, as suggested by Rebesco et al. (2014).

This work is structured as follows: the study area is described in Section 2; data, methodologies, and both numerical and theoretical models are described in Section 3. Analyses and results from numerical simulations of temperature (T) and salinity (S) vertical transects as well as the analysis of the theoretical model for bottom current pathways are in Section 4, for both North Aegean Trough (NAT) and Central Basin. The bottom sediments of the Cyclades Plateau and the Cretan Sea hydrology are analyzed in Section 5. A short synthesis and conclusions are in Section 6.

### 2. Study area

The Aegean basin has a very complex structure of different local environments as canyons, river deltas, coastal or offshore sediments (Maley and Johnson, 1971; Zervakis et al., 2000).
In particular, the NAT consists of the North Sporades (≈ 1460 m deep) and Athos basins (≈ 1150 m deep). At its southern boundary, the Northern Aegean Sea is connected by the Limnos-Sporades channel (≈ 500 m deep at the sill) with the Central Aegean Sea. The North Aegean Sea hydrology is characterized by a dense bottom layer, an upper layer of Levantine Intermediate Waters (LIW) and a surface layer. Often, a much colder and fresher surface Black Sea Water (BSW) crosses the Dardanelles and works as a thermal insulator over the NAT (Plakhin, 1972; Poulos et al., 1997; Zervakis et al., 2000; Gertman et al., 2006).

The North Aegean Sea is characterized by a sea floor with deep trenches, shelves and sills. About its sedimentary patterns, Hamann et al. (2008) focused attention on the sediment origins: they analyzed the different roles of west Anatolian fluvial suspension load, Aeolian dust, the transport of the suspension load to the core position by the anticlockwise surface currents. Karageorgis et al. (2005) investigate, in particular, the effect of Thermaikos Gulf rivers and currents. They find that in that Gulf dense water flows enter from SE and then are trapped in a cyclonic gyre, just off the shelf. Consequently, fluvial sediments discharged from the rivers remain mostly present within the plateau.

The Central Aegean Sea (in the following Central Basin) is divided in the northern Cyclades Plateau and the Skyros, North Ikaria and Chios basins (Fig. 1). Its hydrology is similar to that of the NAT, but with a different inflow of BSW. Indeed, modified BSW enters the Central Basin through its NW flank and then follows the cyclonic flow of the Aegean Sea.

Lykousis (2001) analyzed the subaqueous bed-forms of the Cyclades Plateau and found dunes, sand waves, 2-D mega-ripples, sand ribbons. He also remarked how such bottom structures would imply large bottom water velocities (from ≈40 to 200 cm/s) that, however, were not recorded during his cruise, where the measured water velocity was about 6–10 cm/s.

The South Aegean Sea is composed by the Ikaria basin, the southern shelf of the Cyclades Plateau and the Cretan basin, the largest depression of this region reaching ≈ 2000 m in depth. The southernmost part of the Cretan Sea communicates with the Eastern Mediterranean through the Cretan Arc straits.

3. Methods and data

3.1. In situ hydrographic and seismic reflection data

The few available hydrologic data (i.e. 31 CTD casts) during year 1987 were collected in Aegean Sea during the LIA-5-87 cruise onboard of the R/V Aegeo. The North Aegean Sea was sampled from 27 February to 3 March 1987 after a week of strong cold northerlies (G2T3). The hydrographic casts (Fig. 2) are analyzed in order to characterize the typical hydrological structures and, in particular, to validate the hydrographic information we derive from the numerical model, in terms of temperature, salinity, and potential density.

The mapping of late Quaternary sedimentary facies on the Cyclades Plateau (Fig. 3) is based on the interpretation of extensive 10 in. airgun seismic-reflection profiles collected during years 1986–1992 through several oceanographic cruises with the R/V Aegaeo (Tripsanas et al., 2015).

3.2. Satellite Sea Surface Temperature (SST) data

We analyze an Aegean subset of SST data from the AVHRR-Pathfinder, focusing on those regions characterized by dense water formation. These SST data have a daily resolution of ≈ 4 km (two times per day) and are derived from the 4-km Global Area Coverage (http://www.nodc.noaa.gov).

In particular, we use monthly averaged SST time series from 1981 to 2006 to infer dense water formation processes from SST anomalies. We focus on 1987 (Fig. 4), which is considered as the first year of a cold
period (Zervakis et al., 2000) even though it is not the coldest year of the 1982–2006 time-series.

3.3. The PROTHEUS model

The numerical outputs we used in this study are from the PROTHEUS system (kindly provided by Dr. Sannino), i.e., a coupled model composed by the RegCM3 atmospheric regional model and a MITgcm ocean model (Artale et al., 2010). The coupling of these two models is accomplished by the OASIS3 coupler (Valcke and Redler, 2006; for more details see Appendices A and B).

We stress that, in being a hydrostatic model the PROTHEUS is not particularly suitable for simulating dense water chimney phenomena. Despite this source of uncertainties, we use T and S vertical transects from this model to analyze the main hydrographic setting (and their monthly evolution) in some crucial areas of the Aegean and Crete Seas, an approach that is not largely affected by the non-hydrostaticity of the model.

The PROTHEUS simulation is performed on an area including the entire Mediterranean Sea, where there are also considered inflows from rivers and external seas, as Dardanelles that is here schematized as a runoff. As an example, monthly averages of transect F are shown in Fig. 5A and B, to be compared with the G2T3 measurements in Fig. 2A.

Fig. 3. Shaded bathymetric map of the southwest Aegean Sea, showing (A) the distribution of the bottom-current related seismic facies, and (B, C) examples of moats and sediment drifts in the southwest Aegean Sea.
and B. We moreover note how the PROTHEUS bathymetry is often particularly smoothed in comparison with the real sea bottom (for more details see Appendix B).

The model validation (Appendix B) shows that the PROTHEUS monthly averaged temperatures are about 1–2 °C lower for the surface layer of BSW (simulated as a layer of cold fresh water often less than

---

**Fig. 4.** Monthly satellite data of winter-spring SST over the Aegean Sea during year 1987.
40 m thick, Figs. 2A and 5A). This large discrepancy probably is due to a poor parameterization of the Dardanelles outflow of BSW in the PROTHEUS model and/or a numerical bias (Artale, personal communication). For the other layers less deep than ≈ 500 m this kind of discrepancy is ≈ 0.5 °C. On the other hand, PROTHEUS salinity is about 0.9 psu lower for BSW and then is rather realistic, perhaps a 0.1 psu lower for the largest depths, while the PROTHEUS density is rather realistic on the surface and good for the deep layers.

We remind that those of PROTHEUS are Eulerian data that concern local values in a given month, and thus the Lagrangian evolution of the previous measured values, although fully advected by the Eulerian model, can be affected by uncertainties related to Lagrangian chaos (Palatella et al., 2014). We also point out that the use of the Eulerian velocities from this model, which can be potentially used for our analysis, are not suitable for our goal since bottom numerical currents are usually affected by large errors (due to, e.g., coarse vertical resolution, lack of salinity data for model initialization) and thus might result unrealistic (Sannino and Artale, personal communications).

3.4. Theoretical models

Models for dense water evolution over a slope describe that a dense water, immediately after its formation, can either start a slow geostrophic motion with velocity \( u \) or it can downflow as rather quick a cascade. In this first case the pioneering streamtube model (Smith, 1975; Killworth, 1977, among others) describes the geostrophic motion of a vein of dense water along the slope. The balance of Coriolis force and buoyancy in a stratified fluid gives a dense water moving approximately along the isobaths (Smith, 1975; Killworth, 1977; Estournel et al., 2005).

However, along this motion the vein slightly deepens (a second order effect of friction and entrainment) and its cross section increases. In addition, just under the main streamtube a small parallel flow of slightly denser current often occurs.

In more detail, since the streamtube has some turbulent characteristics due to the bottom friction, its motion can erase adjacent waters, that is, a streamtube thickness \( h \) in the point \( x \) along its path evolves as

\[
\frac{udh}{dx} = E|u-\nu|
\]  

(1)

where \( \nu \) is the ambient water velocity (Turner, 1973) and \( E \) is the entrainment parameter, usually considered as a function of the Froude number \( F_r \). In the Mediterranean Sea \( E \) is \(-10^{-4}\)–\(-10^{-5}\) (Baringer and Price, 1997; Falcini and Salusti, 2015).

Different schematizations, due to Shaw and Csanady (1983) and Shapiro and Hill (1997), correlate the evolution of a dense water initial patch with the bottom density gradients, thus estimating a density current flowing horizontally with speed (Nof, 1983)

\[
u_{sof} = \frac{s g' \rho}{f}
\]  

(2)

where \( s \) is the sea bottom slope, \( g' \) is the reduced gravity and \( f \) is the Coriolis parameter.

This kind of models also allows computing the small dense water deepening. Killworth (2001) applied a Zilitinkevich and Mironov (1996) analysis on meteorological and oceanographic downflows based on a Froude number \( F_r \) characterization, and found that such outflows often follow a simple trajectory characterized by a constant increase of the current depth \( z \), so obtaining \( dz/dx \approx \) constant. This
4. Analyses and results

4.1. The hydrographic data

North of the Samothrace island, the C transect of G2T3 (Fig. 2) – simulated by the F transect of PROTHEUS – is about 50 km at the east of the W transect of G2T3. The core data of the transect C of G2T3 (Fig. 2a and b: \( T \approx 13.50 \, ^\circ C, S \approx 38.75 \) psu, \( \sigma_\theta \approx 29.25 \) at \( z \approx 170 \) to \( 350 \) m), and the more westward transect W (not shown, \( T \approx 12.50 \, ^\circ C, S \approx 38.40-38.75 \) psu, \( \sigma_\theta \approx 29.23 \) at from \( \approx 150 \) to \( 380 \) m) reveal processes of dense water formation over the North Samothrace shelf. G2T3 observed the densest transect \( T \) over the more southern Limnos-Lesbos Plateau (not shown, \( T \approx 13.60-13.80 \, ^\circ C, S \approx 38.95 \) psu, \( \sigma_\theta \approx 29.33 \)).

In these G2T3 sections, the typical hydrological structure of the Aegean Sea is revealed by temperature and salinity (Fig. 2a, b), with relatively high values of \( T \) and \( S \) at intermediate and deep layers. In the surface layer the cold and fresh BSW, often thinner than \( \approx 30-40 \) m, can be seen (Zervakis and Georgopoulos, 2002; Pazi, 2008).

During the same period, Zervakis et al. (2000) synthesized the available marine water densities near Limnos island as \( \sigma_\theta \approx 29.32 \), while further south in the Cyclades, at \( \approx 660 \) m depth, is \( \sigma_\theta \approx 29.18 \); at east of the Island of Crete was found \( \sigma_\theta \approx 29.16 \) at 1570 m depth and \( \sigma_\theta \approx 29.15 \); west of that island, at \( \approx 1270 \) m depth. In general, Zervakis et al. (2004) interpret the deep Cretan Water as a mixing of LIW and dense water of NAT origin, a characterization in agreement with our results.

4.2. SST patterns from remote sensing

In his careful analysis of the Aegean Sea SST, Josey (2003) recognized the essential importance of the thermal effects for the density evolution of the surface waters during the EMT winters, in comparison with evaporation. From January to May of 1987, the coast near Samothrace, the Thermaikos gulf, and the Limnos-Lesbos shelf appear to be particularly affected by strong winds from east, funneled by the Greek and Turkish orography (Sayin and Beşiktepe, 2010 and Sayin et al., 2011). These winds determine lower temperatures in these regions in comparison to Central and Southern Aegean Sea. In turn, during June, the Cyclades Plateau has rather lower temperatures (Fig. 4) than the northern area.

In addition, Fig. 4 shows that the 1987 was not particularly cold, while the Turkish coast seems colder than the other shelves (Zervakis and Georgopoulos, 2002; Sayin and Beşiktepe, 2010 and Sayin et al., 2011). No particular thermal effect is evident over the Cyclades during 1987 while, in the following years, the water temperature decreases by \( \approx 1-1.5 \, ^\circ C \) (see Fig. S1 in Supplementary). This highlights the role of the winter 1986–87 in preconditioning the EMT, then fully developed during the early ‘90s (Sayin et al., 2010 and 2011).

Summarizing, dense water “sources” are recognized in the Samothrace and Limnos-Lesbos shelves, as suggested by G2T3, or in the Cyclades Plateau as well (Zervakis et al., 2000 and 2002; Sayin et al., 2010 and 2011; among others). Although in the Thermaikos Gulf surface dense waters are evident, this gulf is also affected by a remarkable fresh water input from rivers (Estournel et al., 2005).

4.3. Numerical results in the NAT basin [transsects F, R, and G]

In order to simulate lacking field data and to assess dense vein pathways and dynamics, we analyzed monthly averaged \( T \) and \( S \) data in some selected transects as obtained from the PROTHEUS model (Figs. 6, 7, 8; for details see Fig. 1 and Table 1). Results are synthesized in 3 heuristic “monthly T/S plots”, namely an overlap of T/S diagrams relative to the same geographical points but for different months. In these plots we indicate different waters as XX YYY (e.g., F 40 or NB 100, see Table 1), where XX is the name of the transect in Fig. 1 and YYY is the water depth (m).

Fig. 6 shows the PROTHEUS monthly averages of T/S diagrams of NAT during January, February and March of the year 1987. We focus on the transects shown in Fig. 1 and Table 1:

i) the Samothrace shelf transect (F);
ii) the shelf and the deepest part of Thermaikos gulf (R);
iii) the Limnos-Lesbos shelf (G).

During January 1987 the densest and coldest waters are found in the Samothrace shelf (F 40), with a PROTHEUS abundant \( \sigma_\theta \approx 29.65 \), where both \( S \) and \( \sigma_\theta \) decrease considerably during February and March, likely due to a BSW interaction (Fig. 6). This salinity decrease is a general effect since it also occurs in the other superficial (G 40) and (R 40) points, which become also remarkably cold. Either a meteorological interaction or an interaction with BSW are likely the main causes of such a dense water formation.

A strong cooling also occurs in the Limnos shelf (G 40), where the densest water were measured by G2T3. PROTHEUS data in January show \( T \approx 14.2 \, ^\circ C \) and in March \( T \approx 12.3 \, ^\circ C \) for the (G 40) water, a cooling likely influenced by particularly dramatic storms during March 1987 (Lagouvardos et al., 1998).

Temperature in the deepest point of the Samothrace shelf (i.e., F 470), as well as for the Thermaikos gulf (i.e., R 760, R 360), remains rather constant while salinity increases from January to March (Fig. 6). Therefore, there is no evidence of any mixing between this deep (F 470) water with the overlying shelf (F 40), although some mixing of...
(F 470) is possible with (G 40), i.e. the I transect of G2T3, or with the sill water (G 200).

About the Limnos-Sporades channel, the point (G 200) "moves" towards — appears to be mixed with — the denser (G 40) water. Indeed (G 200) increases its potential density of \( \rho \approx 0.1 \) in February \( \rho \approx 0.2 \) in March, finally reaching \( \rho \approx 29.42 \) in April. This agrees with the hypothesis of a robust interaction between (G 40) and (G 200) waters (Fig. 6).

In the next we will show that also the (F 40) water has a similar interaction with (G 200), but with some months of delay. On the other hand, a mixing of (F 470) with the (G 200) sill water, partially flowing northward to balance the massive southward flow of (G 40) and (G 200) waters, could explain the deep (F 470) time evolution.

In turn, the points R in the Thermaikos Gulf have a rather irregular behavior, probably due to March storms or local river outflows, while only (R 300) looks mainly mixed with (F 40). These mixings support the G2T3 idea that the (F 40) dense water moved westward along the isobaths over the shelf break in a slow geostrophic motion. Of particular interest is that all densities in the southern sector of the basin are smaller than those of (F 40) and (G 40) during January, namely \( \alpha_b \approx 29.65 \) and 29.43 (Fig. 6).

4.4. Numerical results in the Central basin [transects G, NB, H, and D]

We now analyze the following PROTHEUS monthly averages of T/S diagrams of the Central basin during January, February and March of the year 1987 (Fig. 7) in:

i) the (G 200) and (G 40), already discussed above;
ii) the southern boundary of the Central Basin, i.e. transect H with its eastern point at 40 m depth (H 40), near Bodrum (Fig. 1) and its sill (H 250);
iii) the point (D 200) near the island of Ikaria, East of the Cycladic shelf;
iv) the NB transect (part of transect B from north of the Cyclades till Lesbos), i.e. points (NB 50) East of Lesbos, (NB 100) near the island of Andros, at the northern Cycladic border, and the deep sill (NB 500).

Fig. 7 shows that in these months the near-surface points (G 40) and (H 40) become remarkably colder and fresher, mainly in March, likely due to the strong front over the Aegean Sea (Lagouvardos et al., 1998). The deep (H 250), (NB 500), and (D 200) have a common evolution, i.e. a rather steady T and an increasing S from January to March, indicating a mixing with the salty, deep current influencing the northern Cycladic border, and the deep sill (NB 500).
Table 1
Characteristics of the PROTHEUS transects. In red two points now shown in Fig. 1 (black dots).

<table>
<thead>
<tr>
<th>Transect</th>
<th>Orientation</th>
<th>Origin point</th>
<th>End point</th>
<th>Length (Km)</th>
<th>Bottom slope</th>
<th>Critical points and depths (m)</th>
<th>Point symbol</th>
</tr>
</thead>
<tbody>
<tr>
<td>F</td>
<td>N-S</td>
<td>25°11'E 40°58'N</td>
<td>25°11'E 40°06'N</td>
<td>94</td>
<td>$5 \times 10^{-3}$</td>
<td>Northern shelf, 40 Bottom, 470</td>
<td>F 40</td>
</tr>
<tr>
<td>R</td>
<td>NW-SE</td>
<td>23°06'E 39°56'N</td>
<td>24°14'E 39°27'N</td>
<td>115</td>
<td>$7 \times 10^{-3}$</td>
<td>Northern shelf, 40 Intermediate, 300 Bottom, 760 Sporades, 360</td>
<td>R 40 R 300 R 760 R 360</td>
</tr>
<tr>
<td>G</td>
<td>E-W</td>
<td>25°33'E 39°60'N</td>
<td>24°13'E 39°21'N</td>
<td>165</td>
<td>$8 \times 10^{-3}$</td>
<td>Limnos, 30 sill, 200</td>
<td>G 40 G 200</td>
</tr>
<tr>
<td>NB</td>
<td>NE-SW</td>
<td>25°42'E 39°26'N</td>
<td>24°53'E 38°00'N</td>
<td>232</td>
<td>$5 \times 10^{-3}$</td>
<td>Lesbos shelf, 50 deep sill, 500 Cyclades shelf, 100 Turkish shelf, 20 deep Ikaria, 250</td>
<td>NB 50 NB 100 NB 500</td>
</tr>
<tr>
<td>H</td>
<td>E-W</td>
<td>27°34'E 37°11'N</td>
<td>25°16'E 37°11'N</td>
<td>204</td>
<td>$2 \times 10^{-3}$</td>
<td>Cyclades plateau, 100 South Ikaria, 200 Cyclades, 1000 deep Cretan, 2000</td>
<td>H 20 H 250 H 100</td>
</tr>
<tr>
<td>D</td>
<td>N-S</td>
<td>25°05'E 37°13'N</td>
<td>27°04'E 35°01'N</td>
<td>340</td>
<td>$4 \times 10^{-3}$</td>
<td>Southern Cyclades, 800 western Cretan, 1500</td>
<td>D 200 D 1000 D 2000 SB 800 SB 1500</td>
</tr>
<tr>
<td>SB</td>
<td>NE-SW</td>
<td>24°25'E 37°08'N</td>
<td>22°53'E 35°05'N</td>
<td>357</td>
<td>$6 \times 10^{-3}$</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Fig. 9. Pathways of dense water masses in the context of the streamtube (see text). It also indicated the position of points (NB 50), (NB 100) and (NB 500). Green arrows represent the evolution of the (F 40) water; yellow arrows represent the evolution of the dense (G 40)–(G 200) water; red arrows represent the northward evolution of the (H 250), (NB 500), and (D 200) deep waters.
has a slightly varying T
PROTHEUS T/S data (transects
the (\(\sigma_\theta\)) values in NB at different depth (50, 100 and 500 m) in SI.

<table>
<thead>
<tr>
<th>Months</th>
<th>T</th>
<th>S</th>
<th>(\sigma_\theta)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>NB50</td>
<td>NB100</td>
<td>NB500</td>
</tr>
<tr>
<td>January</td>
<td>14.51</td>
<td>14.53</td>
<td>13.21</td>
</tr>
<tr>
<td>February</td>
<td>13.79</td>
<td>14.20</td>
<td>13.26</td>
</tr>
<tr>
<td>March</td>
<td>13.20</td>
<td>13.02</td>
<td>13.40</td>
</tr>
<tr>
<td>April</td>
<td>13.80</td>
<td>13.49</td>
<td>13.51</td>
</tr>
<tr>
<td>May</td>
<td>13.70</td>
<td>13.35</td>
<td>13.36</td>
</tr>
<tr>
<td>June</td>
<td>14.28</td>
<td>13.01</td>
<td>12.98</td>
</tr>
</tbody>
</table>

\(\sigma_\theta\) \approx 29.37 \text{ to } \approx 29.08 \text{ in June during the same time interval. By considering that the deep (NB 500) water at the sill had a density } \approx 29.24 \text{ during January–March, we can argue that this water probably mixes with (G 200) and reaches } \psi_F \approx 29.34 \text{ in May; } \psi_F \text{ increases to } \approx 29.36 \text{ in June, very similar to the } \psi_F \approx 29.4 \text{ observed by G2T3. }

The possible candidates for the origin of such density changes are those water coming from the Limnos-Sporades sill (G 200) and shelf (G 40), or a deep northward current along the Turkish coast of salty (H 250) water (Fig. 1), or just cold storms. In this regard, the (H 250) sill water evolution shows that its salinity indeed increases from \(\approx 38.72 \text{ psu in January to } \approx 38.86 \text{ psu in April and May, and then in June it decreases to } S \approx 38.78 \text{ psu, somehow similar to the other southern (D 200) deep point. This implies a decrease of the deep sea salty northward transient in June. An effect of this (H 250) water in the Cyclades (NB 100) water mass looks difficult, since this water is warmer and saltier than the (NB 100) water.

The remarkable density increase during May–June of (NB 500) is likely due to the overlapping of all these cold flows: first from (G 200) water, mostly in May, then for the summer an interaction in June–July with dense waters of (G 40) origin, while no direct effect of storms on the (NB 500) water is particularly evident. In synthesis, deep (NB 500 m) water in the Cyclades plateau appears to be mixed mainly with cold water from NAT and Limnos-Lesbos Plateau, as suggested in the early studies of Zervakis et al. (2000, Zervakis and Georgopoulos, 2002).

4.5. Numerical results in the Cretan basin [transects SB and D]

About the onset of the main EMT outflow from the Cretan Sea, PROTHEUS T/S data (transects SB and D) show that the point (SB 800) has a slightly varying T \(\approx 14.4–14.6 \text{ °C and } S \approx 38.82–38.86 \text{ psu and the } \psi_F \approx 29.04–29.08. \) For (D 2000; in the Cretan Basin, Table 1) the salinity is around 38.73 psu, T \(\approx 13.55 \text{ °C and the density } \approx 29.06–29.08. \) For the western (SB 1500), the temperature is T \(\approx 13.74 \text{ °C while } S \approx 38.74–38.76 \text{ psu and the density } \approx 29.17 \text{ always with very small variations}. \) This suggests that deep layers south of the Cyclades Plateau and in the deep Cretan Sea both salinity and temperature show little variations, and are similar to field data quoted in Zervakis et al. (2000). From these critical points we infer that no outflow of the Aegean dense water is evident during the January–June 1987 period.

4.6. Velocities and pathways from the theoretical models

About the marine water dynamics in the NAT, we notice that the (G 40) densest water (that formed south of Limnos) moves geostrophically along the isobaths and, through the time, reaches similar T–S values of the (G 200) in-channel water (Figs. 1 and 6). For the (F 40) water, the data agree with the streamtube dynamics: a near geostrophic pathway along the NAT shelf-break (Figs. 6 and 9). This can explain the observed lack of mixing between (F 40) and (F 470). This westward motion along the isobaths receives also some support from the dynamic topography observed by Gertman et al. (2006) in the northernmost part of the NAT during 1988. Therefore, it is reasonable to assume that the (F 40)

<table>
<thead>
<tr>
<th>Table 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Estimated values of the streamtube model velocities along transects in SI.</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Transect</th>
<th>(g)</th>
<th>(s)</th>
<th>(h)</th>
<th>Not velocity</th>
<th>Killworth velocity</th>
</tr>
</thead>
<tbody>
<tr>
<td>F</td>
<td>(10^{-3})</td>
<td>(5 \times 10^{-3})</td>
<td>80</td>
<td>0.08</td>
<td>0.3</td>
</tr>
<tr>
<td>R</td>
<td>(8 \times 10^{-4})</td>
<td>(7 \times 10^{-3})</td>
<td>60</td>
<td>0.06</td>
<td>0.2</td>
</tr>
<tr>
<td>G</td>
<td>(10^{-3})</td>
<td>(8 \times 10^{-3})</td>
<td>70</td>
<td>0.08</td>
<td>0.2</td>
</tr>
<tr>
<td>NB</td>
<td>(10^{-3})</td>
<td>(5 \times 10^{-3})</td>
<td>100</td>
<td>0.05</td>
<td>0.3</td>
</tr>
<tr>
<td>H</td>
<td>(5 \times 10^{-4})</td>
<td>(2 \times 10^{-3})</td>
<td>70</td>
<td>0.01</td>
<td>0.2</td>
</tr>
<tr>
<td>D</td>
<td>(5 \times 10^{-4})</td>
<td>(4 \times 10^{-3})</td>
<td>50</td>
<td>0.02</td>
<td>0.2</td>
</tr>
<tr>
<td>SB</td>
<td>(10^{-3})</td>
<td>(6 \times 10^{-3})</td>
<td>50</td>
<td>0.06</td>
<td>0.2</td>
</tr>
</tbody>
</table>

By considering the distance from the shelf where (F 40) to the Limnos-Sporades sill (i.e., \(\approx 400 \text{ km}) and a water velocity of \(\approx 4 \text{ cm/s}, \) as estimated by G2T3, it would take \(\approx 120 \text{ days (Figs. 1 and 6) to cover such a distance. Similar velocities can be estimated over the Samothrace shelf from applying the Nof velocity in Eq. (1). Along this hypothetical pathway (Figs. 6 and 9) the mean bottom slope in between the Athos and Sporades sector (i.e., a distance of \(\approx 100 \text{ km}) would give a delay of about 90 days, implying a March–April dense water arrival in the Limnos-Sporades channel. Velocities estimated by using the Killworth model (Eq. (1)), which are not affected by the local bottom characteristics, are much larger and reach values of \(\approx 30 \text{ cm/s). \) This would imply an earlier arrival of dense (F 40) water at the Limnos-Sporades channel during February (Table 3).

In addition, the application of the streamtube model (Eq. (2)) to the Samothrace transect C of G2T3 (i.e., transect F of the PROTHEUS transects) gives a dense water westward velocity \(\approx 4–5 \text{ cm/s). Considering the small velocities of these dense water streamtubes in the NAT, their thickness } h \text{ is expected to increase as } dh/dx \approx E/2 \approx 10^{-5}. \text{ We remark that these time estimates, however, are affected by significant uncertainties due to the presence of several local, topographic effects that cannot be easily considered here. }

From Eq. (1), we also obtain that the thickness } h \text{ of the } (F 40) \text{ water layer increases as } dh/dx \approx E/2 \text{ along its } \approx 400 \text{ km long pathway in the NAT shelf-break. By assuming a constant E/2, we obtain that entrainment gives an increase of } h \text{ around } 15 \text{ m at the end of its travel. This small variation is neglected in the following discussion, although some are considerable and not-captured mixing processes (e.g., confluence of different water masses towards a similar pathway) is certainly present. Finally, from the streamtube model we estimate a deepening of this } (F 40) \text{ cascading of } \approx 1 \text{ m/km. }

Regarding the Central basin, T/S diagrams suggested a northward, high salinity flow along the Turkish coast. The (G 200) water is about 230 km (NB 500; Fig. 1) and, by considering a shelf slope of \(\approx 8 \times 10^{-3}, \) a local \(u_{Nof} \approx 8 \text{ cm/s (see also Lykoumis, 2001)) would give one month of delay from the } (G 200) \text{ density peak in April (Figs. 7, 8, and Table 3), giving a first cold water arrival in } (NB 500) \text{ May. From the Limnos shelf, at } \approx 480 \text{ km from the } (NB 500).
500) point with a $u_{\text{max}} \approx 5 \text{ cm/s}$, one has about 3 months of delay and thus a dense water arrival can be estimated in June–July (Fig. 10).

4.7. Geophysical results

The upper 200 m of the sedimentary packet in the Cyclades Plateau most probably is the product of late Quaternary sedimentary processes (Anastasakis et al., 2006). It consists of the six seismic facies shown in Fig. 3. Parallel, continuous, mid to low-amplitude seismic reflections represent hemipelagic sediments. In turn high to mid-amplitude, discontinuous to incoherent reflections and overlying on a hard substrate without further sound penetration, are usually interpreted as shallow marine and fluvial sediments. Lenses of incoherent mid- to low-amplitude reflections interbedded in hemipelagic sediment represent mass-transport deposits (MTDs). Seismic facies related to bottom-current are expressed by erosional channel-like features, which are bounded on their sides by mounded seismic packets, consisting of high-amplitude incoherent to semi-parallel, discontinuous mid- to low-amplitude seismic reflections. Similar seismic facies have been described from many ocean environments around the world (Rebesco et al., 2014, and references therein) and are usually interpreted as moats and sediment drifts. These sedimentary structures mark the main entrance of the deep water we investigate here between the Islands of Euboea and Andros (Channel B in Fig. 3). Moreover, they occur along some interesting pathways towards the Myrtoon Basin and the Cretan Sea.

5. The Cyclades plateau deposits

We now relate the above results on dense water hydrodynamics and pathways with the evidence of bottom sediments in the Cyclades Plateau (Fig. 3). According to the distribution of the sediment drifts, the bottom currents that affect their formation processes enter the Cyclades Plateau through a strait in the North-East sector of this basin (i.e., between the Islands of Euboea and Andros, NB 100 point in Figs. 1 and 9) and exit towards west and southwest through three deep straits in the Myrtoon Basin and the Cretan Sea. This flow is depicted by the sedimentary structures (sand waves, dunes, and sand ribbons) already identified by Lykousis (2001) in the same area. The continuation of the sediment drifts at the much larger water depth of the Myrtoon Basin indicates that the bottom currents exhibited a strong cascading feature, suggesting that they have large densities.

We notice that bottom velocities, as obtained from the streamtube model, are rather small (Table 2) with respect to those that are needed to form such sedimentary structures (Rebesco et al., 2014). However, this region is characterized by a very complicated bottom topography that have a remarkable influence on the bottom water velocity and their temporal evolution. It could indeed happen that impulsive strong currents, affected by irregular bottom constrictions, give origin to strong nonlinear phenomena. Indeed, Lykousis (2001) from the analysis of such sedimentary structures predicted velocities up to 1–2 m/s while his own measurements were 7–10 cm/s, in some agreement with our estimated velocities (Table 3). Moreover, streamtube models deal with steady motions, while the Lykousis (2001) bottom structures probably captured only the paroxysmical stage of the currents, which might not have lasted for a long time.

Finally, we stress that in the streamtube case lower velocities are expected because we focus our analysis just on a single year, prior to the EMT event. Moreover, Lykousis (2001) estimates would be, in turn, rather similar to the Killworth model velocities. Therefore, these sedimentary results somehow support the model here proposed which, accordingly, agrees with the southward dense water outflow off the Cyclades Plateau. There, one can indeed remark at the boundary of the Cyclades Plateau and the Myrtoon basin (Fig. 3) some evidence of bottom current dynamics, a particular, interesting feature that suggests future investigations.

6. Conclusions

In this paper, we analyze the thermodynamics of Aegean Sea currents during the year 1987, just before the onset of the EMT event (Roether et al., 2013).
The complexity of the EMT event has been largely analyzed by several authors. G2T3 found very dense water masses and assumed that the densest waters of the Cretan Sea arrived via slope convection from the Cycladic Plateau shelf, although specific dense water paths were not identified.

By combining satellite, numerical, field, and theoretical information, we here complement and what was found in seminal work (e.g., Zervakis et al., 2000; Theocharis et al., 1999a, 1999b; Lascaratos et al., 1999; Malanotte-Rizzoli et al., 1999; Nittis et al., 2003) by presenting a novel viewpoint about the Aegean Sea dense water dynamics. Our analysis considers the northern dense water formation zones observed by Zervakis et al. (2000) in the light of a stream tube model (Smith, 1975; Killworth, 1977).

From SST satellite data, we recognized the main sources of very dense water in the Samothrace shelf and the Limnos-Lesbos Plateau. From T and S data, simulated by the PROTHEUS numerical model, we found that the densest water from the Samothrace shelf flowed geostrophically along the NAT isobaths, as a streamtube, towards the Limnos-Sporades channel. There, it mixed with a dense water from the Limnos-Lesbos Plateau and moved geostropically towards the Cyclades during April, May and June with different pathways and arrival times. In addition, it is reasonable to argue that the heavy storm that occurred on March 1987 had a significant role on this dynamics. A different high-salinity, northward flow along the Turkish coast, probably related to wind induced upwelling (Sayin, 2010 and 2011), seemed to play a secondary role. Our stream tube approach essentially supports the Zervakis et al. (2000) analysis.

Our findings were then related to the presence of sedimentary structures in the area. Results suggested that the main entrance of dense water in the Cyclades Plateau is between the Islands of Euboea and Andros. Moats and sediment drifts in this area agree with this southward, with CTD data for the North Samothrace transect (Artale et al., 2010) in the version natural boundary conditions for salinity are used, that is P = R – E (precipitation plus runoff minus evaporation) is treated as a real fresh water flux. Monthly river discharges computed from the RegCM3 total runoff are used by the oceanic model. Catchment basins for 147 rivers falling into the Mediterranean Sea have been reconstructed using Total Runoff Integrated (TRIP) database. Rivers discharging in the Black Sea have been collected to dense fresh water fluxes that reach the Mediterranean Sea via the Dardanelles Strait. A monthly value for the discharge of Black Sea is then included as a further river. The initial condition for the oceanic run is obtained from a previous run performed using a 3D relaxation towards the MEDATLAS-II climatology.

Appendix B. The PROTHEUS model validation

As a check of the numerical simulation over the Aegean Sea, we compared monthly SSTs produced by PROTHEUS with Optimally Interpolated SST (OISST, Marullo et al., 2007) from Pathfinder 5.0 SST data. The agreement between the two time series (Fig. S2) confirms that PROTHEUS is able to reproduce reasonably well the annual SST cycles, though some minor discrepancies between the model and the data do not exceed 1.5 °C and less than 1 °C in average. The difference between OISST and field data shows no time drift (Marullo et al., 2007), as also happens between the model and OISST data and we therefore deduce that the model has no remarkable drift in the overlapping period.

We moreover compare the PROTHEUS hydrologic transect (Fig. 40) with CTD data for the North Samothrace transect (Fig. 2) of G2T3 during March 1987. There is a low surface temperature T ≈ 11.20–11.50 °C (in PROTHEUS simulation T ≈ 9.5 °C), most probably due to a thin surface layer of BSW ≈ 40 m thick. Just below one has T ≈ 12.5 °C while in the deepest layers T reach 13.5 °C (PROTHEUS has T ≈ 12.6 °C down to 100 m, remaining rather homogeneous with 12.8 °C at the sea bottom). As remarked by a referee it is possible that the BSW from Dardanelles Strait is underestimated till a factor 2 by the PROTHEUS model.

About other parts of the Aegean in the same period, Zervakis et al. (2000) synthesize the available field data at Limnos (near Samothrace) with T ≈ 13.30 °C (T ≈ 12.8 °C in the PROTHEUS data) while more South in the Cyclades at 660 m depth T ≈ 14.25 °C (T ≈ 13.4 °C in the PROTHEUS data), at east of the Island of Crete T ≈ 14.05 °C at 1570 m depth (T ≈ 13.6 °C in the PROTHEUS data) and T ≈ 14.10 °C West of that Island, at just 1270 m depth (T ≈ 13.8 °C in the PROTHEUS data). In synthesis the PROTHEUS simulation is about 2 °C low for the BSW, rather poorly simulated, and ≈ 0.5 °C low for the other layers.
In the same F transect, salinity $S \approx 38.00$ psu in the BSF surface layer, rapidly increasing with depth to $38.25–38.50$ psu till reaching about $38.80$ psu in the deepest layers ($\text{PROTHEUS}$ simulates the BSF layer as low as $S \approx 7.0$ psu on the surface, then $S \approx 38.5$ psu at $60–100$ m depth, to a rather homogeneous value $S \approx 38.75$ psu below $200$ m). About other parts of the Aegean in the same period, Zervakis et al. (2000) synthesizes the available field salinity at Limnos (near Samothrace) with $S \approx 38.84$ psu ($S \approx 38.82$ psu in the PROTHEUS data) while more South in the Cyclades at $660$ m depth $S \approx 38.94$ psu ($S \approx 38.95$ psu in the PROTHEUS data), at east of the Island of Crete $S \approx 38.84$ psu at $1570$ m depth ($S \approx 38.8$ psu in the PROTHEUS data) and $S \approx 38.85$ psu West of that Island, at just $1270$ m depth ($S \approx 38.78$ psu in the PROTHEUS data). Again, there is a PROTHEUS salinity about $1$ psu low for BSF and then rather realistic values, perhaps a $0.1$ psu low for the largest depths.

The surface BSF density is less than $\approx 29.05$ while the PROTHEUS density is $\approx 29.0$ in the BSF layer.

References


Baringer, M.O.N., Price, J.F., 1997. Momentum and energy balance of the Mediterranean layer, rapidly increasing with depth to $38.25$ psu and $S \approx 0.1$ psu low for the largest depths.


Baringer, M.O., Price, J.F., 1997. Momentum and energy balance of the Mediterranean layer, rapidly increasing with depth to $38.25$ psu and $S \approx 0.1$ psu low for the largest depths.


