Simultaneous deep and intermediate depth convection in the Northern Levantine Sea, winter 1992

Deep water formation
Intermediate water formation
Convection
Northern Levantine
Winter 1992

Formation des eaux profondes
Formation des eaux intermédiaires
Convection
Nord de la Mer Levantine
Hiver 1992

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ABSTRACT

The northern Levantine Sea is the primary source region for the Levantine Intermediate Water (LIW) in the Mediterranean Sea. The Deep Water (DW) of the Eastern Mediterranean mainly originates in the Adriatic basin, but local contributions from the Levantine Sea have also been suspected in the past. Observations in the northern Levantine Sea during March 1992 shed new light on the above processes, showing simultaneous formation of DW in the cyclonic Rhodes Gyre (Rhodes Gyre) area, and of LIW in the adjacent regions. The deep convection region coincides with the permanent dome structure of the Rhodes Gyre, where overturning of the water is generated by cooling during sufficiently severe winters. The LIW is produced in a much larger area of the northern Levantine Sea than previously thought, by direct surface cooling and mixing of the near-surface stratified waters.

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RÉSUMÉ

Convection profonde et intermédiaire observées simultanément dans le nord de la Mer Levantine durant l'hiver 1992

Le bassin nord de la Mer Levantine est la source principale d'Eau Intermédiaire Levantine (EIL) pour la Mer Méditerranée. L'eau profonde (EP) de la Méditerranée orientale provient principalement du bassin adriatique, mais les contributions locales de la Mer Levantine ont aussi été soupçonnées par le passé. Les observations effectuées dans le nord de la Mer Levantine pendant le mois de mars 1992 éclairent d'un jour nouveau les processus caractéristiques de ces régions-sources. Elles montrent la formation simultanée de l'EP dans l'aire du tourbillon cyclonique de Rhodes (tourbillon de Rhodes) et de l'EIL dans les régions voisines. La région de convection profonde coïncide avec la structure permanente en dôme du tourbillon de Rhodes, où la plongée de l'eau est engendrée par le refroidissement en surface pendant les hivers assez rudes. L'EIL est produite dans le nord du bassin levantin, dans une aire plus large que celle que nous imaginions, par le refroidissement direct de la surface et par le mélange des eaux stratifiées près de la surface.

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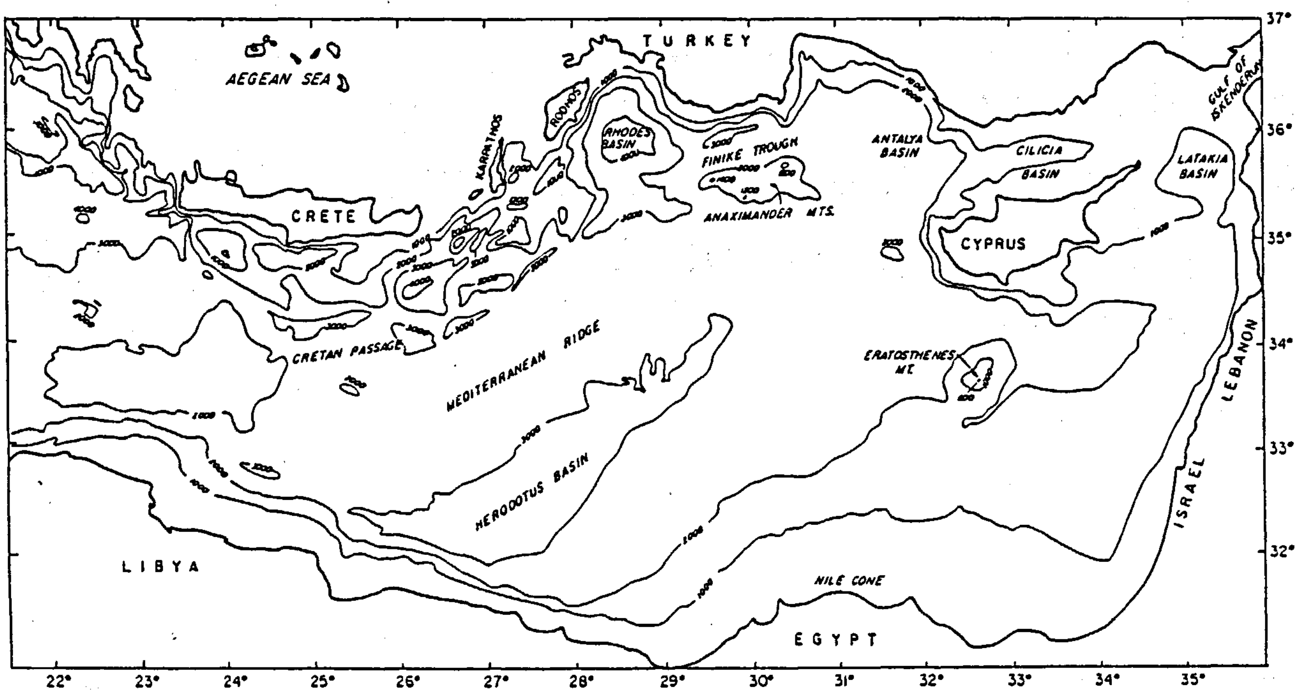

Figure 1

Location map and bathymetry of the Levantine basin (after Özsoy et al.*, 1991).*

INTRODUCTION

The Levantine basin (the eastern part of the Eastern Mediterranean, Fig. 1) circulation, described by a new synthesis (Özsoy *et al.*, 1989; 1991; 1992; Robinson *et al.*, 1991 *a*; 1992) based on recently collected data, consists of a series of dynamically interacting sub-basin scale circulation cells (the Rhodes, Mersa Matruh and Shikmona Gyres) and embedded coherent structures (the Anaximander, Antalya, Shikmona and Southwest Cretan Eddies), fed by bifurcating jet flows (the Central Levantine Basin Current, the Asia Minor Current). Long-term (seasonal) stability is a unique characteristic of the Eastern Mediterranean eddies, and seems to be reproduced in model studies (Robinson *et al.*, 1991 *b*; 1992). It is also known that the basin general circulation, water masses and the distribution of coherent eddies are subject to significant interannual variability (Özsoy *et al.*, 1991, 1992).

The cyclonic Rhodes Gyre (Rhodes Gyre) is a permanent member of the Levantine basin circulation, with a cold dome hydrographic structure (Anati, 1984) revealed in the ocean climatology (Ovchinnikov, 1966; Tziperman and Malanotte-Rizzoli, 1991) of the region. Surface divergence (Ovchinnikov, 1966) near the eastern part of the Cretan arc, associated with local maxima of the wind stress curl (Malanotte-Rizzoli and Bergamasco, 1991; Pinardi and Navarra, 1992) and heat fluxes (Malanotte-Rizzoli and Bergamasco, 1991), as well as the local topography of the Rhodes Basin (Özsoy *et al.*, 1989; 1991; 1992; Robinson *et al.*, 1991 *a*; 1992) are known to be factors of primary importance in driving the gyre. Model studies (Milliff and Robinson, 1992) reveal a coherent current system and short-term oscillations (a few weeks) driven by bursts of baroclinic instability, although the rate of such baroclinic energy conversions does not seem to influence the long-term persistence (Özsoy *et al.*, 1989; 1991; 1992; Robinson *et al.*, 1991 *a*; 1992; Anati,

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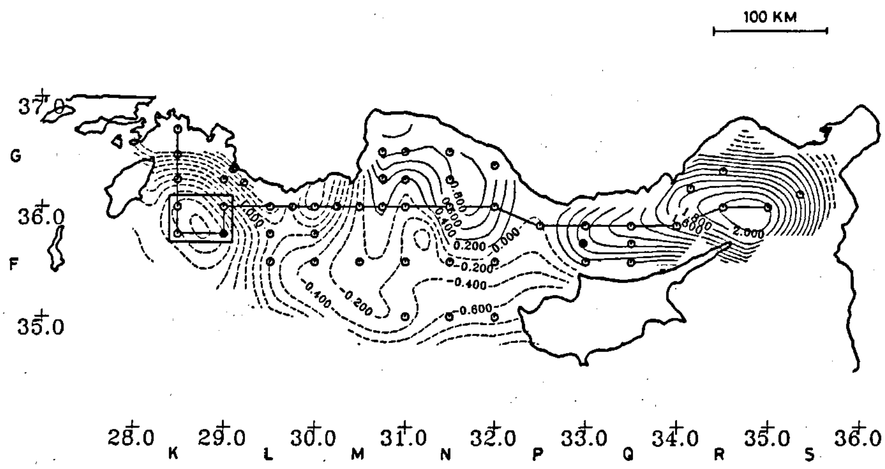
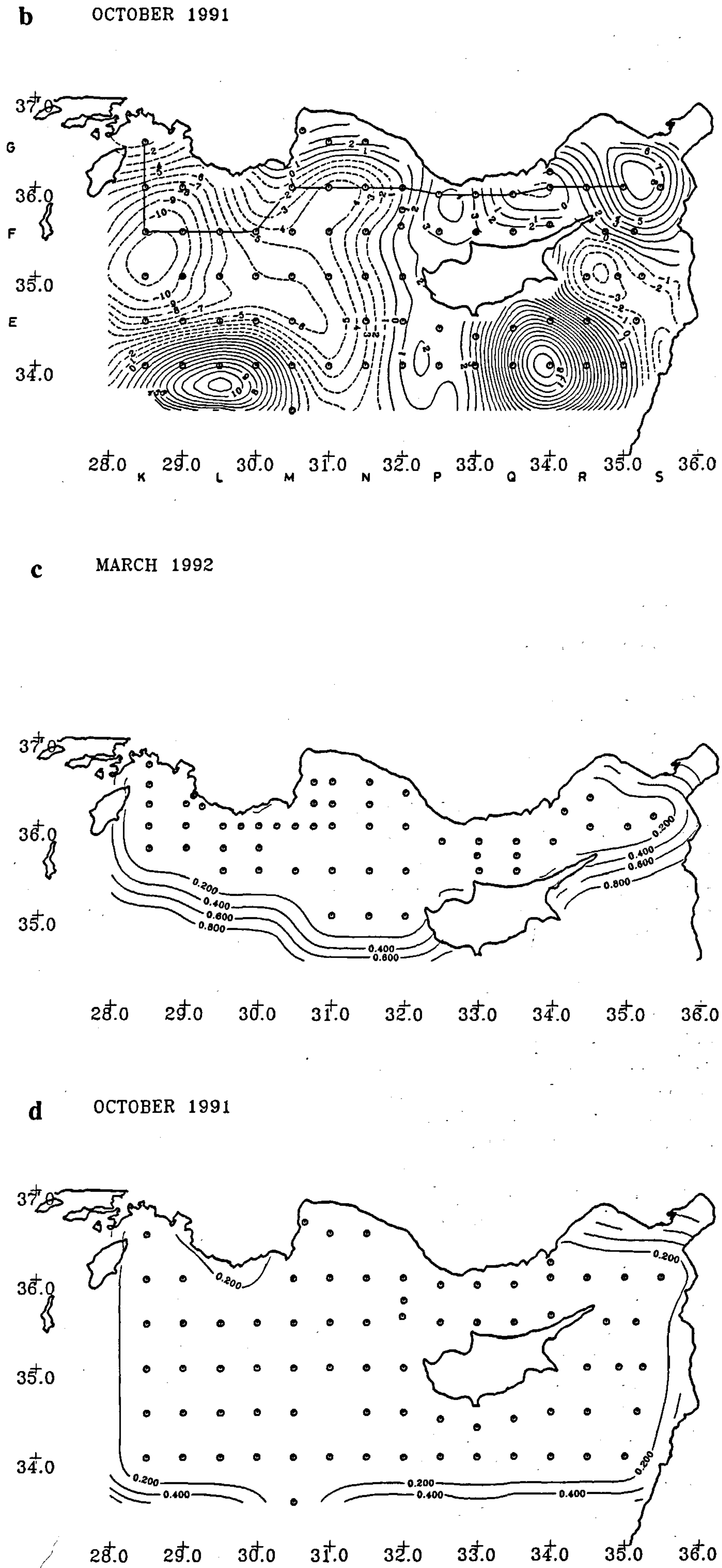
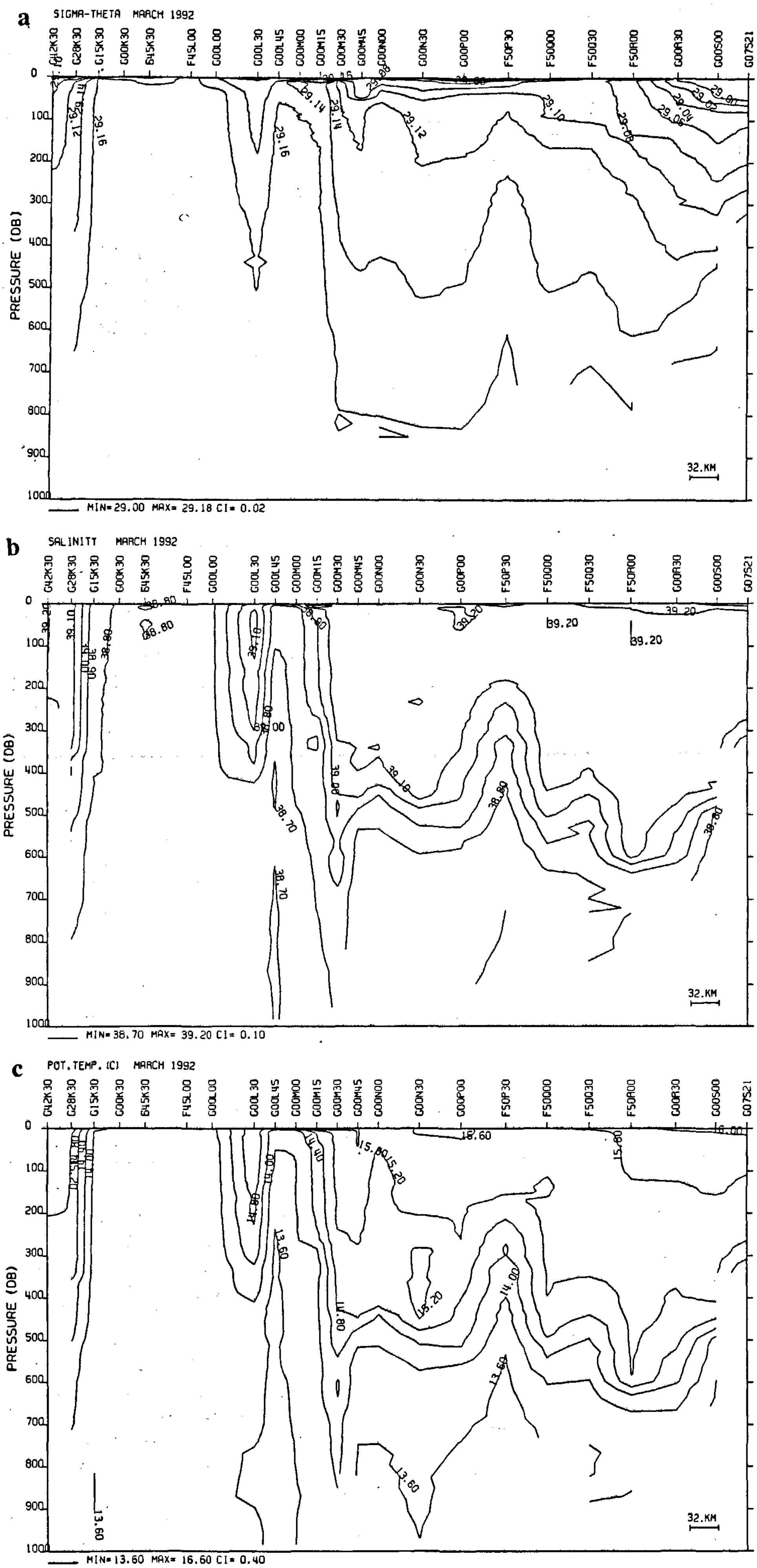




⟸ Figure 2 ⟹

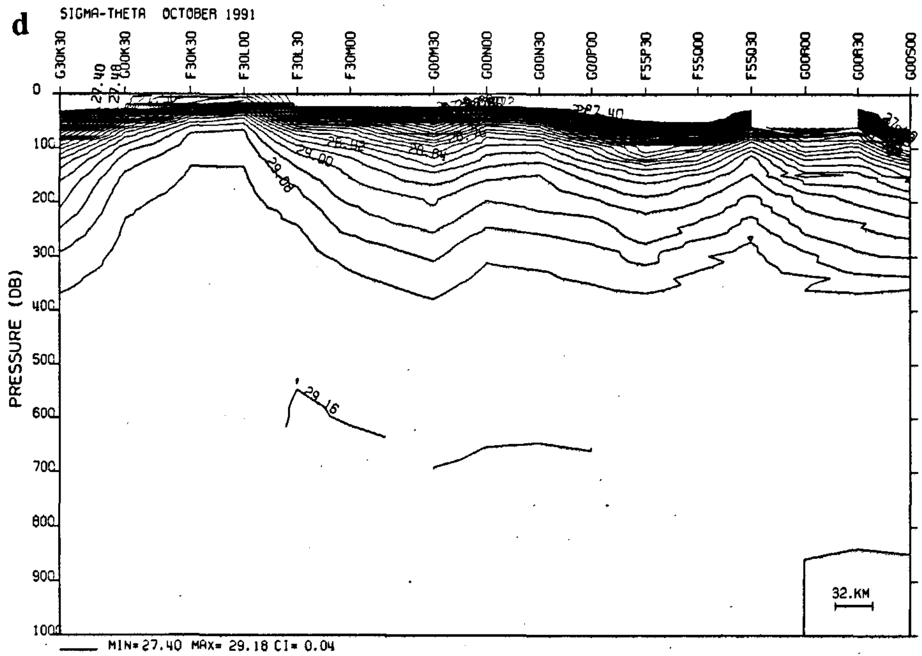
Positions of hydrographic stations and the surface geopotential height referenced to 1000 decibar level of no motion during: a) March 1992; and b) October 1991, and the normalized error variance maps for the objective analyses during: c) March 1992; and d) October 1991. Dynamic topography contours are given in centimetre units, and have different spacing in the two surveys. Solid lines mark the transects depicted in Figures 3 a-f*, in which station positions are identified with letter coding of latitude and longitude shown in Figures 2* a, b. *The four stations in the square domain marked in Figure 2* a *are plotted with different symbols in Figure 4. Closed symbols show the stations for which profiles are given in Figure 5.*



⟸ Figure 3 ⟹

a) Density (σ_θ); b) salinity; and c) potential temperature (θ) distribution along transect of Figure 2 a during March 1992; and d) Density (σ_θ); e) salinity; and f) potential temperature (θ) distribution along transect of Figure 2 b in October 1991.

d



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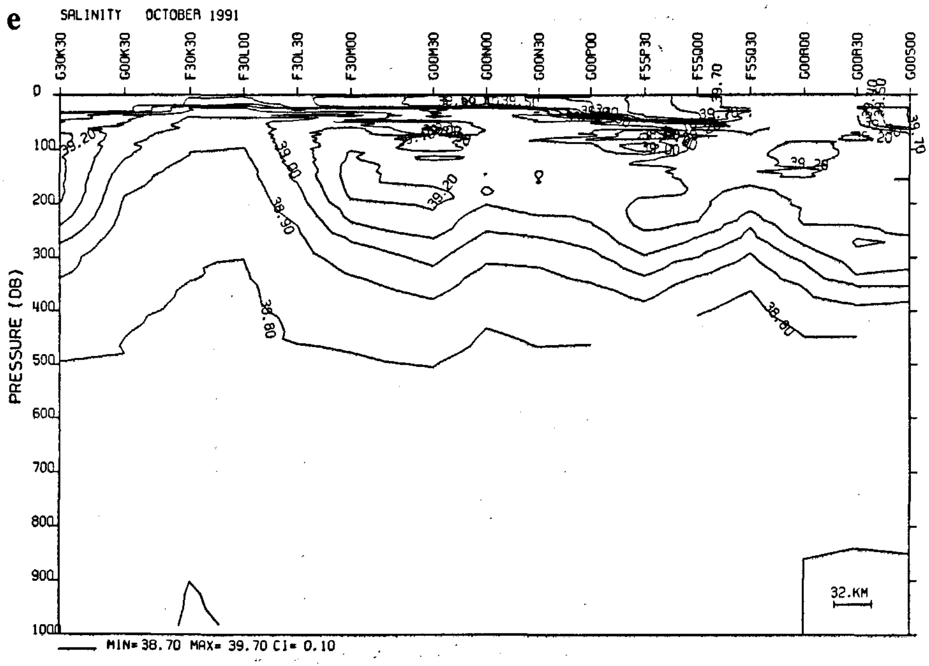

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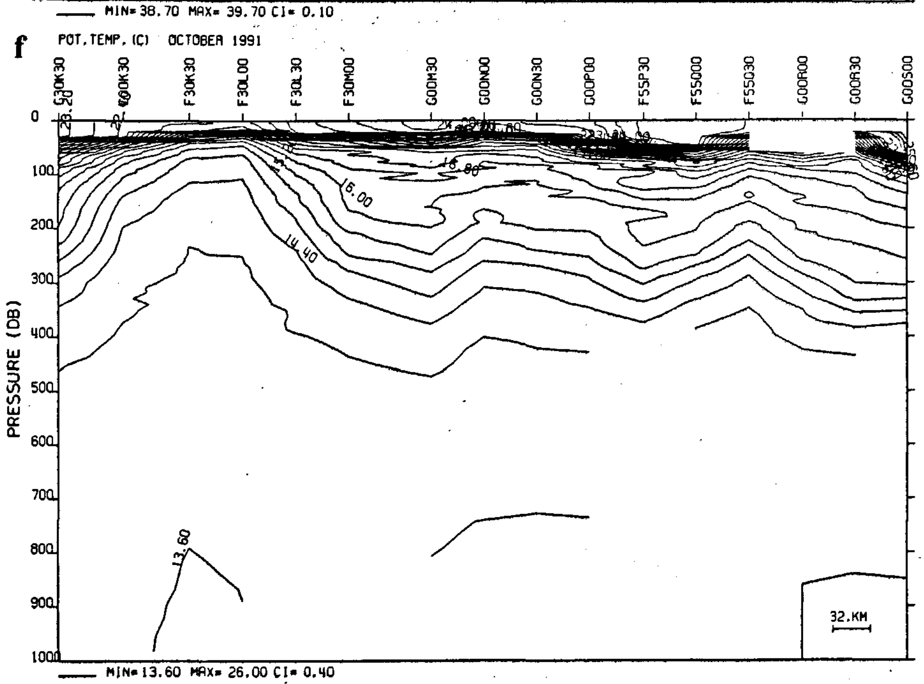

1984) of the gyre, whose size is an order of magnitude larger than the baroclinic Rossby radius of deformation.

The main water types (Hecht *et al.*, 1988) seasonally maintained in the circulation system are the Levantine Surface Water (LSW), identified with the mixed layer in summer, the Atlantic Water (AW), reaching the Levant from its origin in the Atlantic Ocean, the Levantine Intermediate Water (LIW), which is locally produced in the northern Levantine basin, the Deep Water (DW) and the Bottom Water (BW).

The Deep Water (DW) is only a modified form of the Bottom Water (BW) [Roether and Schlitzer, 1991; Schlitzer *et al.*, 1991] which sinks and spreads to the entire basin from sources in the Adriatic Basin (Pollak, 1951). The Adriatic Sea origin of the DW below 1200 m is fairly well established by recent studies (Roether and Schlitzer, 1991; Schlitzer *et al.*, 1991). For the water at shallower depths of 500-1200 m, these results do not seem to contradict other local sources suggested in the adjacent Aegean Sea (Miller, 1972), and within the Rhodes Gyre (Ovchinnikov, 1984). Recently, wintertime deep water formation has been shown to occur in the Rhodes Gyre (Gertman *et al.*, 1990).

Although the formation mechanism is not sufficiently understood, the LIW can be found with distinct, yet variable characteristics in the entire northern Levantine source region, in pools trapped within anticyclonic eddies and along the Anatolian coast (Özsoy *et al.*, 1989; 1991; 1992; Brenner *et al.*, 1991; Özturgut, 1976). LIW is characterized by high salinity (39.1), and temperature (15.5°C) values which are attained from the mixture of different water types. The following models have been considered to explain LIW formation: a) isopycnal sinking into adjacent anticyclonic eddies of dense surface water formed in cyclonic regions, *e. g.* Rhodes

Gyre (Ovchinnikov, 1984; Ovchinnikov and Plakhin, 1984); b) continental shelf/slope convection in the adjacent Aegean Sea (Bruce and Charnock, 1965; Georgopoulos *et al.*, 1989); c) local convective overturning in anticyclonic eddy centres (Brenner *et al.*, 1991); d) Ekman flux accumulation near the coast of water formed by basin-wide wind-induced mixing (Feliks, 1991).

DW FORMATION

The fact that only a single set of direct observations, in March 1987, exists for deep convection in the Rhodes Gyre (Gertman *et al.*, 1990) underscores the possible intermittency of convection episodes, or their interannual recurrence. For example, earlier observations during the winters of 1974 (Özturgut, 1976), 1986 and 1989 (Özsoy *et al.*, 1989; 1991; 1992) did not indicate any signs of deep convection. On the other hand, one of our own cruises, discontinued in February-March 1987 (Özsoy *et al.*, 1991; 1992) due to severe storms, failed to cover the Rhodes Gyre region, and therefore missed the chance for observing the reported deep convection event (Gertman *et al.*, 1990).

We obtained unique observations of deep convection during the cruise of the RV *Bilim* in March 1992. The surface circulation and station positions are shown in Figure 2 *a*. The density (σ_θ), salinity and potential temperature (θ) cross-sections in Figures 3 *a-c*, along a section outlined in Figure 2 *a*, provide evidence for a strong case of mixing in the entire northern Levantine Sea, resulting in a 1000 m (the maximum depth reached by the observations) deep convection area of uniform deep water density in the Rhodes Gyre, and mixing to depths of 500-600 m (below the depth of the thermocline) elsewhere. The surface circulation (Fig. 2 *b*) and density, salinity and potential temperature distribution (Fig. 3 *d-f*) during a preceding survey in October 1991 are also shown for comparison. The normalized error variance maps for the surface circulations are provided in Figures 2 *c*, *d*.

A reference level of 1000 dbar was used in dynamic computations. The dynamic height is calculated in cm units, and the observational mean is subtracted. The objective analyses are made on a regular grid of (1/4)° spacing, using a correlation model fitted to the observations. The gridded data are contoured by the ZCSEG contouring routine of the Plot 88 Library (Plotworks Inc.), with 4 grid subdivisions and 2 subsegments for each linear contour segment. The contouring is masked by coastal boundaries and when the normalized error variance exceeds 0.6.

($\Delta\theta/\Delta S$) standard deviations for the homogeneous stations (G00K30, F45K30, F45L00, G00L00), respectively, are 0.0234/0.0060, 0.0409/0.0076, 0.0052/0.0007,and 0.0245/0.0022. The wide convective patch of homogeneous deep water coincides with the centre of the Rhodes Gyre (Fig. 2 *a*, *b*), and extends to a secondary structure east of the main patch in the Finike trough. The oscillations in the isolines near the secondary feature (Fig. 3 *a-c*) may suggest adjustment processes (Hermann and Owens, 1991), or baroclinic instabilities, observed in similar formations elsewhere (Gascard, 1978; 1991; Madec and Crépon, 1991). For example, the distance between the primary and secondary fronts (~ 100 km) is only larger than 2π times the first baroclinic radius (~ 10 km), and can be interpreted as a measure of the most unstable wavelengths.

The homogeneous waters at the centre of the deep structure are slightly denser than the deep waters in the surrounding area (Fig. 4), showing the role of buoyancy losses to the atmosphere which appear to be sufficient to cause overturning in the water column.

Deep water formation by convective processes is typically characterized by a preconditioning phase, followed by a violent mixing phase (Gascard, 1978; 1991; Killworth, 1979; Madec and Crépon, 1991). The mass of dense fluid resulting from the deep convection comprises a large

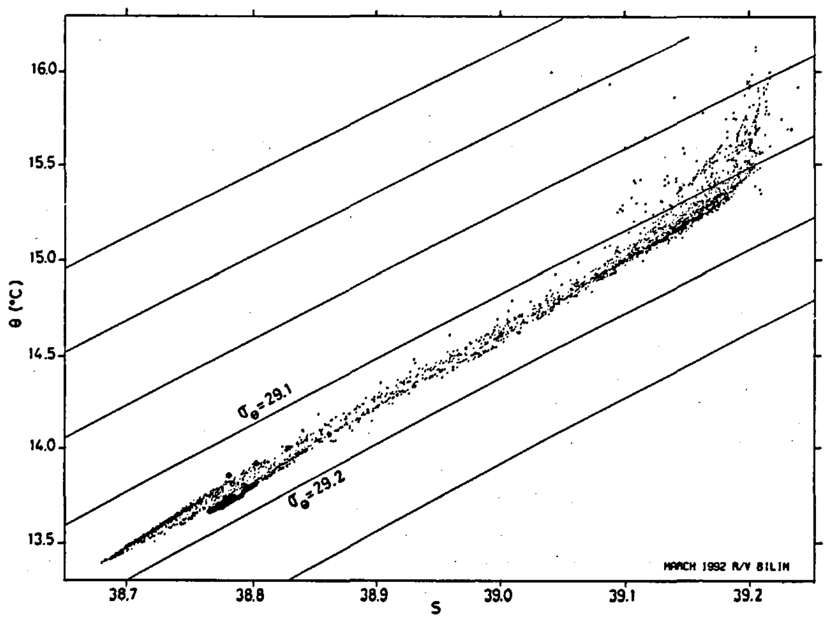

Figure 4

Potential temperature (θ) and salinity (S) in the March 1992 survey of the northern Levantine. The data from four stations in the Rhodes deep convection area (square in Fig. 2 a) are plotted with larger round symbols to differentiate them from the rest of the stations.

reservoir of potential energy, which is converted to kinetic energy by radiating near-inertial waves during the initial geostrophic adjustment (radiation) phase (Hermann and Owens, 1991) and by gravitational and baroclinic adjustment during the baroclinic instability (advection) phase (Gascard, 1991; Madec and Crépon, 1991; Hermann and Owens, 1991); the larger patches (with size several times larger than the Rossby radius, as in the present case) decay mainly by advective processes at Rossby radius scales. The breakup of the dynamic height-based March circulation, as evidenced by the lack of continuity along the coast of its streamlines and the presence of small scale oscillations, differs markedly from the earlier circulation depicted in autumn (Fig. 2 *a*, *b*), and the higher density of the new DW suggests ongoing (Fig. 4) dynamic adjustment processes at the time of the March observations.

The circulation estimates depend on the assumed depth of no motion, so that the amplitude of the barotropic component remains uncertain; in reality it can be larger (Gascard, 1978) in comparison with the baroclinic components computed by the analyses. Note that the estimated velocities in the March 1992 circulation are much smaller than those in October 1991 (the contour intervals are different in Fig. 2 *a* and 2 *b*); this provides further evidence that the flow is not dynamically adjusted in the March 1992 case: either the assumption of geostrophy used in dynamical height calculations is not justified, or the flow has not been able to recuperate from the mixing events which have destroyed the pre-existing stratification. A similar situation was observed in the northern Levantine circulation during the March 1989 observations (Özsoy *et al.*, 1992).

Coincidence with a permanent cyclonic gyre is a unique characteristic of the Rhodes deep convection region. The only other example in the Mediterranean is the western Mediterranean deep convection, where a sub-basin scale transient cyclonic circulation is generated by Mistral winds (Gascard, 1991; Madec and Crépon, 1991; Leaman and Schott, 1991; Schott and Leaman, 1991). Conversely, the surface buoyancy losses and vertical mass exchange due to DW formation appear responsible for generating cyclonic circulations elsewhere (Crépon *et al.*, 1989, Madec and Crépon, 1991), and recurrent DW formation episodes could be contributing factors to the cyclonic circulation of the Rhodes Gyre.

A weakly stratified water mass below the 50 m thick surface layer capping the Rhodes Gyre dome (Özsoy *et al.*, 1989; 1991; 1992; Anati, 1984) in summer renders it susceptible to cooling induced convection in winter (Fig. 3 *d-f* and 5 *a*). Mesoscale eddies exist in the region (Özsoy *et al.*, 1992; Robinson *et al.*, 1992; Milliff and Robinson, 1992); however, there is no evidence to show that the preconditioning of the subsurface waters occurs through a baroclinic instability mechanism, considered to be an essential element in other cases (Gascard, 1991). On the other hand, the erosion of internal structure from within, by upwelling, can facilitate the preconditioning of the Rhodes Gyre. In most earlier observations (*e. g.* Anati, 1984; Özsoy *et al.*, 1989; 1991; 1992) the Rhodes Gyre is characterized by almost uniform temperature and salinity below a thin surface layer capping its dome structure. This thin surface stratified layer may yield to cooling only during extreme events. For example, in March 1989, the 15°C isotherm became as shallow as 10 m at the centre of the Rhodes Gyre (Özsoy *et al.*, 1992), but apparently the surface cooling was not sufficient to generate overturning in this case. On the other hand, comparison of the two vertical profiles obtained in October 1991 and March 1992 in Figure 5 *a* shows that the mixing exceeded 1000 m. The salinity of the mixture accounts for mixing between the surface and deep waters, and the temperature difference between the two cruises indicates a net heat loss in Figure 5 *a*.

LIW FORMATION

As compared to the deep convection in the Rhodes Gyre centre, the intermediate depth convection in the surrounding waters appear to be limited by the pre-existing stratification. In these regions, cooling seems to reduce the stability of the water column, but is not able to erase it completely, as shown by a finite residual density gradient (compared to uniform density in the Rhodes deep convection area) in Figures 3 *a* and 5 *b*. The combined effects of cooling and wind mixing lead to a vertical redistribution of temperature and salinity evident in Figure 5 *b*, and an efficient mixed layer deepening.

In contrast with the Rhodes Gyre region (Fig. 5 *a*), Figure 5 *b* indicates a larger net loss of buoyancy in the outlying areas where LIW is formed; however, since initially (in October) the water is more saline and warmer than the Rhodes Gyre region, a complete overturning by the loss of static instability is not achieved. Although the final (March) density appears uniform due to the plotting scale in Figure 5 *b*, a finite density stratification survives (*e. g.* Fig. 3 *a*) in these regions. We also note in Figure 5 *b* that the temperature and salinity signals have deepened to about 700 m depth by deep convection. The temperature and salinity make finite jumps at the base of the deep mixed layer, but due to the compensating effects (on density) of temperature by salinity, a linear density stratification is produced (Fig. 3 *a*), without leading to a corresponding finite jump in density.

The thermocline during October 1991 is at depths of 200-300 m (Fig. 3 *f* and 5 *b*), eroding to depths of 500-600 m by deep convection in March 1992. This convective process accounts for LIW formation in the entire northern Levantine excluding the Rhodes Gyre, because the final properties (15-16°C and 39.1-39.2 salinity shown by the linear data cluster in Fig. 4) are typical of LIW.

LIW can be formed by one-dimensional mixed layer development in the Shikmona anticyclonic eddy (Brenner *et al.*, 1991), and along the coastal periphery of the cyclonic circulation of the Asia Minor current (Özsoy *et al.*, 1992), where the pre-existing weak stratification complements convective mixing. LIW formation by such local convection occurred in a wide area of the northern Levantine during March 1989 (Özsoy *et al.*, 1992). Stability dependent mixing driven by recurrent winter

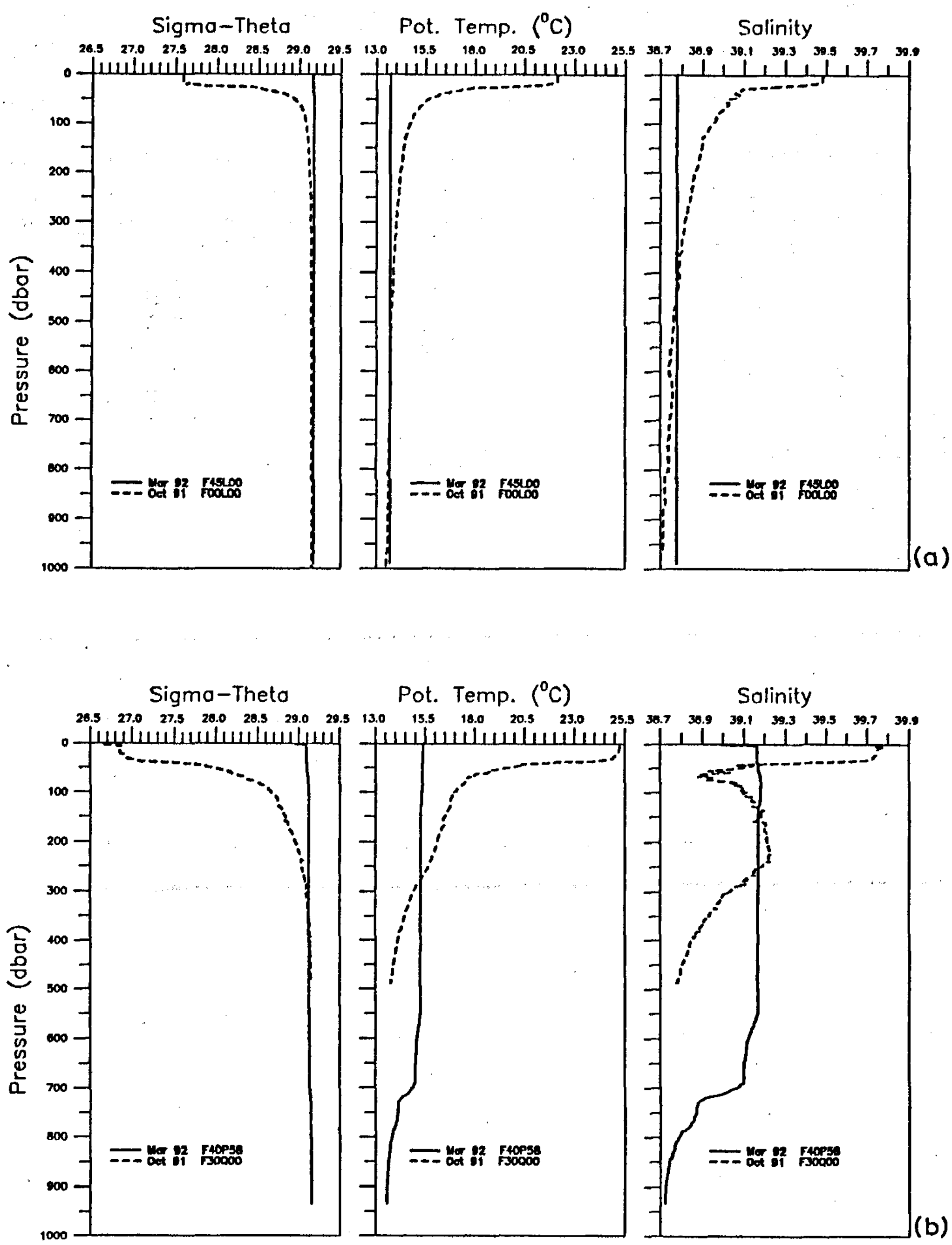

Figure 5

Typical profiles of density (σ_θ), potential temperature (θ) and salinity during October 1991, and density (σ_θ), potential temperature (θ) and salinity during March 1992 observations, at stations (a) within the Rhodes Gyre deep convection region; (b) in the intermediate depth convection region excluding the Rhodes Gyre. The positions of the corresponding stations are shown in Figure 2 a *and* 2 b.

storms (Feliks, 1991) is another valid model able to generate uniform masses of LIW over wide areas of the northern Levantine Sea; the 1992 violent mixing, leading to uniform properties in the entire northern Levantine region, seems to support the predictions of such a model. A notable feature of both the March 1989 and the present observations is the linear θ-S relationship between the LIW and the DW masses (Fig. 4), with almost uniform density of the water column, and diminished abundance of the other near-surface water masses in the entire region.

It seems irrelevant whether isopycnal sinking contributed to the formation of LIW in winter 1992, because efficient vertical mixing is evident throughout the basin, which, however, does not preclude isopycnal sinking at other times (*e. g.* Özturgut, 1976; Ovchinnikov, 1984; and Özsoy *et al.*, 1989), during which high salinity patches have been observed on the periphery of the Rhodes Gyre, and trapped along the coast. Data from the various parts of the region in March 1989 and March 1992 show numerous temperature and salinity inversions, which may be the result of horizontal spreading and active convection.

The impact of mixing on the basin-wide circulation and biological processes

The circulation existing prior to the winter in October 1991 was characterized by the extensive influence of the Rhodes Gyre. Compared to a set of earlier long-term measurements (Özsoy *et al.*, 1989; 1991; 1992), the October 1991 circulation differed considerably. In fact, in the earlier history of the circulation, a qualitative change was detected after 1987, when the originally bifurcating flow (partly

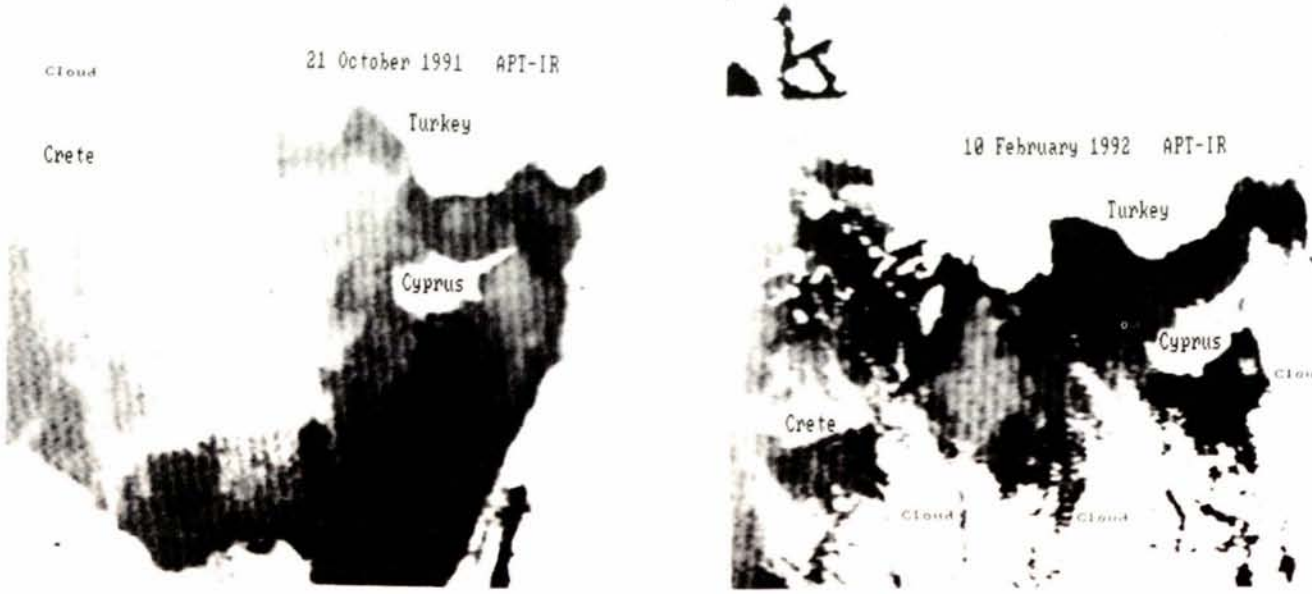

Figure 6

Automatic Picture Transmission (APT) infrared satellite images from: a) 21 October 1991 (NOAA-12); and b) 10 February 1992 (NOAA-10) without atmospheric and geometric corrections. Lighter shades correspond to colder temperatures.

circulating along the Levant coasts and the island of Cyprus, and partly flowing along the rim of the Rhodes Gyre and joining the Anatolian coast) changed to a more coherent flow along the periphery of the Levantine basin. This situation appears to have lasted until August 1990 (Özsoy *et al.*, 1992); in October 1991 it relaxed to a strong circulation around an expanded Rhodes Gyre, and northerly flow joining the Anatolian coast west of Cyprus, and strong anticyclones in the southern part of the domain (Fig. 2 *b*).

Although the factors driving the Levantine basin circulation are quite diverse, it is tempting to note, in retrospect, that the transition to cyclonic peripheral circulation flowing along the basin and turning around Cyprus observed in 1987 could perhaps have a correlation with the influence of the deep water formation in winter 1987, reported by Gertman *et al.* (1990). Similarly, Astraldi and Gasparini (1992) have found that western Mediterranean winter convection in the Ligurian-Provençal basin leads to periodic intrusions of Tyrrhenian water through the Corsican channel. The impact of deep water formation on the larger scale horizontal ocean circulation has been demonstrated by Crépon *et al.* (1989) and Barnier *et al.* (1989).

Satellite images (Fig. 6 *a*, *b*) in the same periods as our observations (October 1991 and February 1992) confirm the above analyses of the oceanographic features. The infrared images were obtained locally from NOAA satellites, in the low resolution APT (automatic picture transmission) format, and therefore were not corrected for atmospheric effects. The light shades only indicate relatively colder water as compared to the darker shades, without any particular scale assigned. In Figure 6 *a*, the Rhodes Gyre is shown to cover a large area between the island of Cyprus and the Cretan archipelago, confirming the circulation system estimated in Figure 2 *b*. The two intense anticyclonic eddies south of the Rhodes Gyre and south of Cyprus, and others near the Gulf of Iskenderun and southeast of Crete (not covered by Fig. 2 *b*) are also indicated as warm patches of water. In Figure 6 *b*, we observe similar features to Figure 2 *a*, a smaller area covered by the Rhodes Gyre, undulations near the Finike trough, and relatively warmer water in the anticyclonic region extending south from Antalya Bay. In Figure 2 *a*, the dimensions of the deep convection region, *i. e.* its extension towards the south could not be determined by the data coverage. In Figure 6 *b*, despite the cloud cover in the south, we observe that the area covered by the Rhodes Gyre centre (the deep convection region) extends further to the south.

The impact of the convection processes on biological production appears to be significant. Under normal conditions, nutrient concentrations are low in the surface waters and chlorophyll *a* in the northern Levantine region typically reaches a maximum in a depth range of 35-120 m with maximum observed values of 0.14 µg/l in summer and 0.42 µg/l in winter (Yilmaz *et al.*, 1992). Preliminary analyses (Dilek Ediger and Ayşen Yilmaz, pers. comm.) indicate that the convection event of March 1992 resulted in an unusually homogeneous vertical distribution of chlorophyll *a*, with relatively high concentrations of 0.5 µg/l in the deep convection region, increasing up to 3 µg/l in the frontal region to its east, in consequence to the burst of phytoplankton.

CONCLUSIONS

The main conclusion to be drawn from the above observations is that the LIW or DW can be formed by different processes at different times, or simultaneously by similar convective processes. During milder winters, the LIW could be formed by isopycnal mixing in the Rhodes cyclonic circulation region and slips into the adjoining anticyclonic eddies, where it becomes stored. Under more extreme conditions, such as during winter 1992, LIW can directly be generated by local mixed layer deepening. Similarly, under sufficiently strong cooling, DW can be formed in the permanently preconditioned waters of the Rhodes Gyre.

The extreme mixing events portrayed in this paper seem to be recurrent on time scales of several years; they should be strongly correlated with the climatic factors on a regional scale. Such interannual variability has recently attracted attention based on studies of the circulation and hydrography of the Eastern Mediterranean. On the other hand, variability on these scales essentially implies strong coupling between the ocean and the atmosphere and an important role played by the internal dynamics of the sea, which can only be understood by the pooling of long-term data and continued studies.

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REFERENCES

Anati D.A. (1984). A dome of cold water in the Levantine basin. *Deep-Sea Res.*, **31**, 1251-1257.

Astraldi M. and G.P. Gasparini (1992). The seasonal characteristics of the circulation in the north Mediterranean basin and their relationship with the atmospheric-climatic conditions. *J. geophys. Res.*, **97**, 9531-9540.

Barnier B., M. Crépon and C. Le Provost (1989). Horizontal ocean circulation forced by deep water formation. Part II: A quasi-geostrophic simulation. *J. phys. Oceanogr.*, **19**, 1781-1793.

Brenner S., Z. Rozentraub, J. Bishop and M. Krom (1991). The mixed-layer/thermocline cycle of a persistent warm core eddy in the eastern Mediterranean. *Dynam. atmos. Oceans*, **15**, 457-476.

Bruce J.G. and H. Charnock (1965). Studies of winter sinking of cold water in the Aegean Sea. *Rapp. P.-v. Réun. Commn int. Mer médit.*, **18**, 773-778.

Crépon M., M. Boukthir, B. Barnier and F. Aikman (1989). Horizontal ocean circulation forced by deep water formation. Part I: An analytical study. *J. phys. Oceanogr.*, **19**, 1781-1793.

Feliks Y. (1991). Downwelling along the northern coast of the eastern Mediterranean. *J. phys. Oceanogr.*, **21**, 511-526.

Gascard J.-C. (1978). Mediterranean deep water formation baroclinic instability and oceanic eddies. *Oceanologica Acta*, **1**, 3, 315-330.

Gascard J.-C. (1991). Open ocean convection and deep water formation revisited in Mediterranean, Labrador, Greenland, and Weddell Seas, in: *Deep Convection and Deep Water Formation in the Oceans*, P.C. Chu and J.-C. Gascard, editors. Elsevier, Amsterdam, 157-182.

Georgopoulos D.A., D.A. Theocharis and G. Zodiatis (1989). Intermediate water formation in the Cretan Sea (South Aegean Sea). *Oceanologica Acta*, **12**, 4, 353-359.

Gertman I.F., I.M. Ovchinnikov and Y.I. Popov (1990). Deep convection in the Levantine Sea. *Rapp. P.-v. Réun. Commn int. Mer médit.*, **32**, 172.

Hecht A., N. Pinardi and A.R. Robinson (1988). Currents, water masses, eddies and jets in the Mediterranean Levantine basin. *J. phys. Oceanogr.*, **18**, 1320- 1353.

Hermann A.J., and W.B. Owens (1991). Modelling the geostrophic adjustment and spreading of waters formed by deep convection, in: *Deep Convection and Deep Water Formation in the Oceans*, P.C. Chu and J.-C. Gascard, editors. Elsevier, Amsterdam, 283-308.

Killworth P.D. (1979). On chimney formation in the ocean. *J. phys. Oceanogr.*, **9**, 531-554.

Leaman K.D. and F. Schott (1991). Hydrographic structure of the convection regime in the Gulf of Lions: winter 1987. *J. phys. Oceanogr.*, **21**, 575-598.

Madec G. and M. Crépon (1991). Thermohaline-driven deep water formation in the Northwestern Mediterranean Sea, in: *Deep Convection and Deep Water Formation in the Oceans*, P. C. Chu and J.-C. Gascard, editors. Elsevier, Amsterdam, 241-265.

Malanotte-Rizzoli P. and A. Bergamasco (1991). The wind and thermally driven circulation of the Eastern Mediterranean Sea. Part II: The baroclinic case, *Dynam. atmos. Oceans*, **15**, 355-419.

Miller A.R. (1972). Speculations concerning bottom circulation in the Mediterranean Sea, in: *The Mediterranean Sea*, D.J. Stanley, editor. Dowden, Hutchinson and Ross, 37-42.

Milliff R.F. and A.R. Robinson (1992). Structure and dynamics of the Rhodes Gyre system and dynamical interpolation for estimates of the mesoscale variability. *J. phys. Oceanogr.*, **22**, 4, 317-337.

Ovchinnikov I.M. (1966). Circulation in the surface and intermediate layers of the Mediterranean. *Oceanology*, **6**, 143-148.

Ovchinnikov I.M. (1984). The formation of intermediate water in the Mediterranean. *Oceanology*, **24**, 168-173.

Ovchinnikov I.M. and A. Plakhin (1984). Formation of the intermediate waters of the Mediterranean Sea in the Rhodes cyclonic gyre. *Oceanology*, **24**, 317-319.

Özsoy E., A. Hecht and Ü. Ünlüata (1989). Circulation and hydrography of the Levantine basin. Results of POEM coordinated experiments 1985-1986. *Prog. Oceanogr.*, **22**, 125-170.

Özsoy E., A. Hecht, Ü. Ünlüata, S. Brenner, T. Oğuz, J. Bishop, M.A. Latif and Z. Rozentraub (1991). A review of the Levantine basin circulation and its variability during 1985-1988. *Dynam. atmos. Oceans*, **15**, 421-456.

Özsoy E., A. Hecht, Ü. Ünlüata, S. Brenner, H. I. Sur, J. Bishop, M.A. Latif, Z. Rozentraub and T. Oguz (1992). A synthesis of the Levantine basin circulation and hydrography, 1985-1990. *Deep-Sea Res.*, in press.

Özturgut E. (1976). The source and spreading of the Levantine Intermediate Water in the Eastern Mediterranean, Saclant ASW Research Center Memorandum SM-92, La Spezia, Italy. 45 pp.

Pinardi N. and A. Navarra (1992). Baroclinic wind adjustment processes in the Mediterranean Sea, submitted to *Deep-Sea Res, special issue* (in press)..

Pollak M.J. (1951). The source of the deep water of the eastern Mediterranean Sea. *J. mar. Res.*, **10**, 128-152.

Robinson A.R., M. Golnaraghi, W.G. Leslie, A. Artegiani, A. Hecht, A. Michelato, E. Sansone, A. Theocharis and Ü. Ünlüata (1991 *a*). The Eastern Mediterranean general circulation: features, structure and variability. *Dynam. atmos. Oceans*, **15**, 215-240.

Robinson A.R., M. Golnaraghi, C.J. Lozano, R. Milliff and E. Özsoy (1991 *b*). Data assimilation in quasigeostrophic model with arbitrary coasts and islands. *23rd International Liege Colloquium on Ocean Hydrodynamics, Modelling the Interaction of the Deep Ocean and the Shelf and Coastal Seas, Liege, Belgium.*

Robinson A.R. *et al.* (The POEM Group) (1992). General circulation of the Eastern Mediterranean. *Earth-Sci. Rev.*, in press.

Roether W. and R. Schlitzer (1991). Eastern Mediterranean deep water renewal on the basis of CFM and tritium data. *Dynam. atmos. Oceans*, **15**, 333-354.

Schlitzer R., W. Roether, H. Oster, H.G. Junghans, M. Hausmann, H. Johannsen and A. Michelato (1991). Chlorofluoromethane and oxygen in the Eastern Mediterranean. *Deep-Sea Res.*, **38**, 1531-1551.

Schott F. and K.D. Leaman (1991). Observation with moored acoustic Doppler current profilers in the convection regime in the Golfe du Lion. *J. phys. Oceanogr.*, **21**, 559-574.

Tziperman E. and P. Malanotte-Rizzoli (1991). The climatological seasonal circulation of the Mediterranean Sea. *J. mar. Res.*, **49**, 411-434.

Yilmaz A., D. Ediger, Ö. Baştürk, and S. Tuğrul (1992). Phytoplankton fluorescence and deep chlorophyll maxima in the Northeastern Mediterranean. *Oceanologica Acta*, submitted.