Ocean-atmosphere interaction
Mediterranean Sea
Hydrology
Seasonal variation
Interannual variation

Interaction océan-atmosphère
Mer Méditerranée
Hydrologie
Variations saisonnières
Variations inter-annuelles

Response of the Mediterranean to the water and energy fluxes across its surface, on seasonal and interannual scales

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ABSTRACT

The meteorological conditions prevailing over the Mediterranean, a concentration basin, transform the inflowing Atlantic water into an outflowing, typically Mediterranean, water; the basin is a "small-scale ocean" for which the effects of surface transfers can be evaluated fairly easily.

Knowing the water deficit at the surface and its effects on the sea allows a number of characteristic times to be estimated; the longest one is the residence time of the Atlantic water in the whole sea volume: it is of the order of one century.

This sea is subject to important vertical movements, in particular in certain limited areas in which deep waters are formed in winter. These waters are typical of this basin. A number of mechanisms of this formation are reviewed. The deep water has characteristics which vary slightly for successive winters.

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RÉSUMÉ

Réponse de la Méditerranée aux flux d'eau et d'énergie à travers sa surface, aux échelles saisonnière et inter-annuelle.

Les conditions météorologiques régnant sur la Méditerranée, bassin de concentration, transforment l'eau Atlantique qui y entre en une eau sortante typiquement méditerranéenne; le bassin est « un modèle réduit d'océan », où l'effet des transferts superficiels peut être appréhendé assez aisément.

L'évaluation du déficit d'eau superficiel et de son effet sur la mer permet la mise en évidence de plusieurs temps caractéristiques, dont le plus long est le temps de résidence de l'eau Atlantique dans tout le volume de la mer, qui est de l'ordre du siècle.

Cette mer est le siège de mouvements verticaux importants, en particulier dans certaines aires limitées où se forment des eaux profondes en hiver, qui sont typiques du bassin. Quelques mécanismes de cette formation sont sommairement présentés. Mais les « eaux profondes » ont des caractères qui varient légèrement d'un hiver à l'autre.

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The substance of this paper was presented by H. Lacombe, as an invited review paper on the Mediterranean, during the symposium UGGI 13 (Ocean and atmospheric boundary layers) held at the General Assembly of the International Union of Geodesy and Geophysics in Canberra (Australia), 3-15 december, 1979.

INTRODUCTION

As is well known, the Mediterranean is a system which transforms the Atlantic water entering through the Strait of Gibraltar ($S=36.15^0/_{00}$) into an outflowing, typically Mediterranean water, of much higher salinity ($S \cong 38.4^0/_{00}$) and density, resulting from the influence of the Mediterranean climate on the entering water.

CHARACTERISTIC TIMES

A number of characteristic times to define the rate of this transformatior can be considered.

The "residence time"

The "residence time" τ of the water in the Mediterranean is defined as the ratio of the total volume of the sea to the incoming flux per unit-time, assuming a stationary state. The budgets of water and salt of the sea, assuming that level and salinity are stationary, as has been the case since the beginning of the century, are balanced.

If S_i, V_i, and S_o, V_o, are the mean salinity, and mean volume of the incoming water (i) and of the outflowing water (o) in Gibraltar, the conservation of water and salt, putting: $\Delta V = V_i - V_o = E - P$; $\Delta S = S_o - S_i$, E evaporation, P precipitation and river discharge, gives:

$$\frac{V_i}{S_o} = \frac{V_o}{S_i} = \frac{E-P}{S_o - S_i} = \frac{\Delta V}{\Delta S} \qquad (1)$$

neglecting the effect of small density differences.

S_i and S_o can be determined by hydrocasts in the strait. However, S_o varies widely within the Strait of Gibraltar itself. Reasonable values are $S_i = 36.15^0/_{00}$; $S_o = 37.9^0/_{00}$, for the western part of the strait [near 5°55'W, point "A4" (Lacombe, 1971)].

Given these values, the order of magnitude of τ can be estimated by three ways:

- the direct measurements of V_i or V_o (also near 5°55'W);
- the evaluation of the mean value of $(E-P)$ over the sea;
- the conservation of salt in the whole sea.

All these methods are questionable, for the results depend on parameters which can be only roughly estimated and on few measurements of great practical difficulty.

1) The first way rests upon the consideration of continuity of *water* and *salt* in the sea, on the basis of measurements of the mean salinity and of the outflowing volume of water, in the strait of Gibraltar only.

Direct measurements of current of long duration at several levels in the strait near 5°55'W in the period May-September, in 1958, 1960 and 1961, have led to $V_0 = 36{,}0 \times 10^{12}$ m^3/year (Lacombe, 1971) from which we draw, using the relations (1) and the given values of salinities, $V_i = 37{,}7 \times 10^{12}$ m^3/year (Lacombe, Tchernia, 1972*a*).

Taking into account the inflow of Black Sea water into the Mediterranean (about 0.4×10^{12} m^3/year), the total inflow into the Mediterranean proper is 38.1×10^{12} m^3/year. As the volume of the sea is about 3.71×10^{15} m^3, the residence time is $3.71 \times 10^{15}/38.1 \times 10^{12} \cong 97$ years. The net water gain in Gibraltar is 1.7×10^{12} m^3/year; the net gain of the Mediterranean proper through the Bosporus is about 0.200×10^{12} m^3/year (Tixeront, 1970). The total deficit of the sea surface is then about 1 900 km^3/year, corresponding to a water loss deficit E − P of 75 cm/year. Using Tixeront's (1970) values for P, leads to E = 130 cm/year.

2) The second way (Béthoux, 1977; 1979) is based on the *heat* and *water* budget of the whole sea and on the *continuity of salt*; by calculating the evaporation from the heat budget and by estimating the precipitation over the sea and the river discharge, Béthoux (1977) gives a mean unit area E − P of about 1 m/year, so that the difference $V_i - V_o = 2.5 \times 10^{12}$ m^3/year. From S_o and S_i evaluations and from equation 1, he obtains:

$V_o = 50.5 \times 10^{12}$ m^3/year,

$V_i = 53.0 \times 10^{12}$ m^3/year.

Then the residence time is $3.71 \times 10^{15}/53.0 \times 10^{12} = 70$ years.

3) A third way, even coarser than the former ones, is based on conservation of salt in the whole basin assumed closed in Gibraltar and filled initially with Atlantic water of salinity $36.15^0/_{00}$. The yearly unit area water loss across the surface is assumed to be 1 m. As the *mean* depth of the basin is 1 470 m, a yearly loss of 1 m of water results in a mean increase of salinity of $1/1\,470 \times 36.15^0/_{00} = 0.0246^0/_{00}$/year. However the mean present salinity of the Mediterranean is about $38.6^0/_{00}$ (Miller, Stanley, 1965), and exceeds by $2.45^0/_{00}$ the initial salinity. The time needed for the mean salinity to increase by $2.45^0/_{00}$ under the effect of the water loss across the surface is $2.45/0.0246 \cong 100$ years. In the meantime the level will have dropped by 100 m

Although it is practically impossible to advance an accuracy in the determination of τ by these very different ways, it is encouraging that they all lead to the same order of magnitude for the residence time: one century ($\pm 30\%$).

This is the time-scale to consider for a climate variation to impose an hydrological modification of the whole Mediterranean.

The seasonal time-scale

The above-mentioned evaluation assumes a stationary mean state; however, seasonal variations exist. The annual cycle effectively consists of a large variation of the sea-surface temperature (variation of about 11° corresponding to a σ_t variation of about 4 units); the salinity cycle is weak. However, the density stratification in the sea is controlled, during the warm season (May-October), by the thermal structure (thermocline at about 30 m), while during the cold season (November-April), it is mainly controlled by the salinity structure (pycnocline at a depth of the order of 100 m). $E-P$ is assumed to be 1 m/year per unit area, in both cases.

1) In the first case, $H=30$ m. The yearly increase of salinity over 30 m for a unit area $E-P=1$ m is $\delta s_s = S_i/H = 1.205^0/_{00}$. As, in the Easternmost part of the sea, the summer surface salinity S_s is about $39.3^0/_{00}$ the actual travel time is:

$$\tau_s = \frac{S_s - S_i}{\delta s_s} = 2.6 \text{ years.}$$

2) In the second case $H \cong 100$ m. The yearly increase of salinity over 100 m, for a unit area $E-P=1$ m, is $\delta s_w = S_i/H = 0.361\,5^0/_{00}$. As in the Easternmost sea, the winter surface salinity is about $S_w = 39^0/_{00}$, the actual travel time is:

$$\tau_w = \frac{S_w - S_i}{\delta s_w} = 7.9 \text{ years.}$$

So, the real "age" (since its entering the Gibraltar strait) of the "Atlantic" water present near the surface should be of the order of a few years (3-8 years) in the Eastern Mediterranean, and any parcel of surface water experiences a number of seasonal cycles before reaching the Eastern limit of the sea.

Such approaches are very crude for they assume that the depth of the pycnocline is everywhere the same, in each of the seasons considered; they neglect the spatial variations of the fluxes across the sea surface, the horizontal exchanges generated by the spatial depth variations of the pycnocline and by external factors (winds, ...); they also assume that no exchange of salt takes place across the pycnocline, between the "surface" and "deep" layers.

THE VERTICAL EXCHANGES

However, these vertical exchanges are important at all seasons even in summer in a number of basins. For instance, in the Alboran Sea, while the waters exchanged at the Gibraltar strait (following Béthoux's figures, 1977; 1979) are 53 and 50.5×10^{12} m^3/year, continuity of water and salt in that sea requires that the vertical flux in the area be, according to Béthoux's (1980) calculations: 14.3×10^{12} m^3/year downward, and 13.0×10^{12} m^3/year upward; this corresponds to vertical velocities of the order of 270 m/year in both directions. The strong velocity shear existing in the Alboran Sea, between the upper ans lower layers, may explain the importance of the exchange.

Such calculations cannot be made for all areas in the sea, due to the lack of data. But Béthoux (1980) has made a number of them.

The study of vertical exchanges is clearly very important since, in the Gibraltar strait, the superposition of the inflowing and outflowing waters implies vertical motions in the Sea. As Stommel *et al.* (1973) have shown that the deep water of the Alboran Sea can pass through Gibraltar, these exchanges are likely to affect a great height there.

Finding the areas and times when mixing over a wide range of depths occurs is a most important problem for the Mediterranean.

It is likely that these exchanges mostly take place in the cold season. For the Medoc area (41-43°N; 4-6°E), Seung (1979; 1980), approached the problem along the following lines, on the seasonal time-scale, using Bryden's (1976) method for estimating the low frequency component of vertical velocity.

Under the assumption of geostrophic balance, measuring the currents at two depths, together with the density profile, allows, through the use of the continuity equation, to obtain the vertical velocity between the two depths. Measurements took place from May to November, 1976 and from February to August, 1977. It appears from the results that, after filtering the relatively high frequency motions (>0.01 cph), the mean vertical advection (Fig. 1) is dominantly upwards between February and May, 1977, with velocities of the order of a few millimetres per second. This result is in agreement with those found by Minas (1970) studying the primary productivity. Instead, in summertime (June-October, 1976), the mean vertical motion is very small. Thus, in wintertime, no mean downward velocity has been experienced in the 1977 mild winter (February-April, 1977). It should be concluded that the local deep water formation should occur during downward velocity events of short duration. Such may be the strong, downward, transitory velocity peaks observed during the Medoc cruises (Stommel *et al.*, 1972; Gascard, 1973; 1977; 1978; and see next chapter). These short downward peaks do not practically affect the "mean" ascending winter velocities as found by Seung (1979).

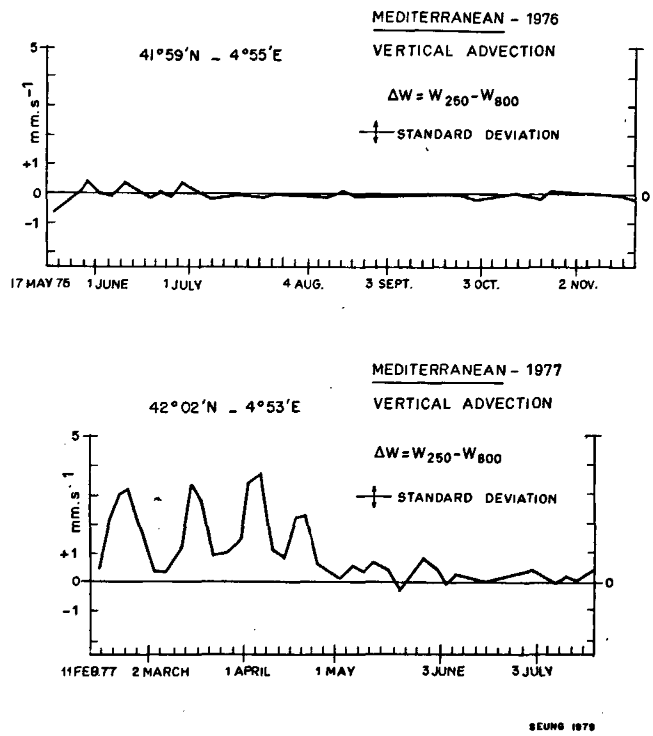

Figure 1

Vertical advection near Bouée-Laboratoire Borha II [May-November 1976 and February-July 1977 (Seung, 1979)].

Advection verticale près de la Bouée-Laboratoire Borha II [mai-novembre 1976 et février-juillet 1977 (Seung, 1979)].

THE FORMATION OF DEEP WATER IN THE MEDOC AREA

The sequence of phenomena

The Mediterranean, North of a line Balearic Islands-Strait of Bonifacio, is in general occupied by a three-layer system (Medoc Group, 1970): a surface layer about 100 m thick marked by a relatively low salinity and deriving from the Atlantic water cyclonic path described in the West-Mediterranean basin by the water entering the Mediterranean in the upper layers through the Gibraltar strait. Below, an intermediate layer of "Intermediate water" (IW), relatively salty and warm; it derives from the "Levantine Water" (LW) formed in winter in a surface layer of about 150-200 m in thickness in the SE-Aegean Sea and the NE-Levantine Basin. The IW occupies a range of depth of 100 to 500-600 m; it follows a cyclonic path also. In deeper layers (500-600 m to the bottom at 2300-2600 m), a Deep Water layer (DW), relatively very homogeneous of about $\theta = 12.7°C$; $S = 38.40\text{-}41\,‰$; $\sigma_\theta = 29.105$.

The presence of a cyclonic circulation around a central area, which occupies an elongated elliptic area from about 41°N-3°E to 43°N-8°E, results in a doming of the

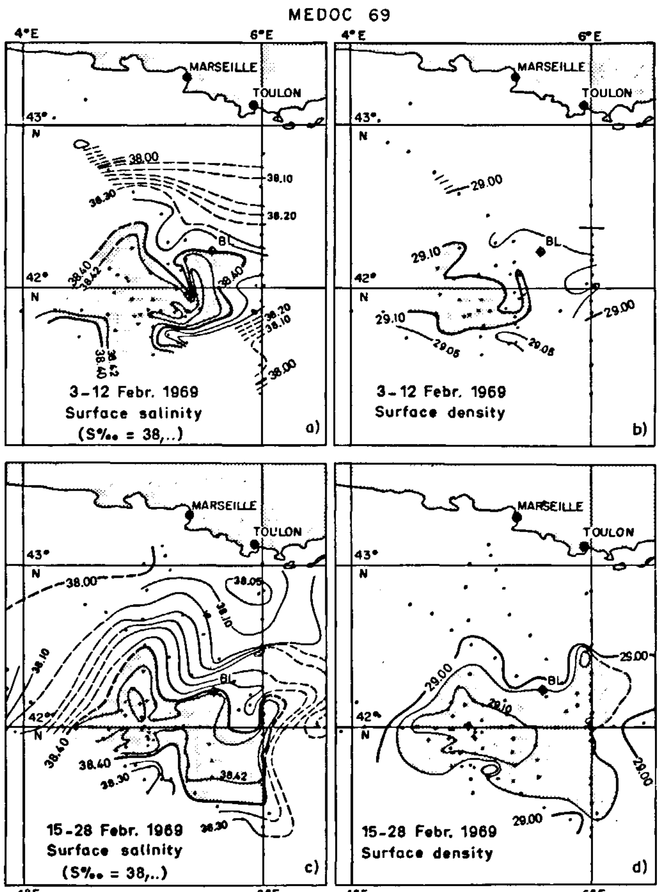

Figure 2

Medoc 1969 cruise. Evolution of the "dense patch" between the beginning and the end of February 1969. Surface salinities and densities (Tchernia, Fieux, 1971). B.L. = Bouée-Laboratoire.

Campagne Medoc 1969. Évolution de la « tache dense » entre le début et la fin de février 1969. Salinités et densités de surface (Tchernia, Fieux, 1971). B.L. = Bouée-Laboratoire.

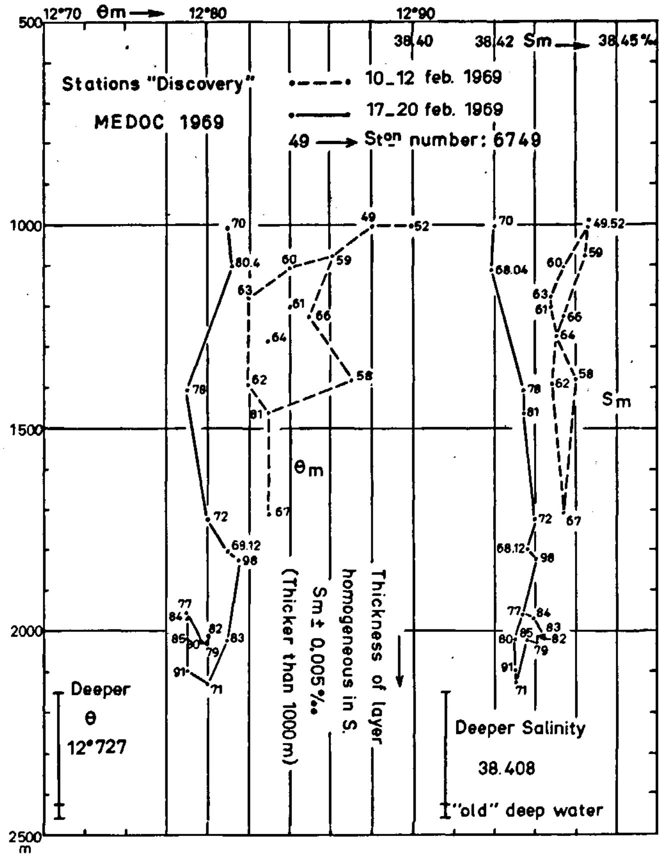

Figure 3

Evolution between February 10-12 and 17-20, 1969 of the mean temperatures and salinities within layers, homogeneous in salinity ($\pm 0.005\,‰$), as a function of their thickness (at least equal to 1000 m from the surface) at "Discovery" stations 1969.

Évolution entre les 10-12 février et 17-20 février 1969 des températures et salinités moyennes au sein des couches homogènes en salinité ($\pm 0{,}005\,‰$) en fonction de leur épaisseur (au moins égale à 1000 m à partir de la surface) en des stations « Discovery » 1969.

σ_θ lines, which are shallower in the center than in the periphery of the gyre.

From the survey made by HMS "Hydra" in January, 1969 it appears (Swallow, Caston, 1973; Sankey, 1973) that the top of the intermediate layer was particulary shallow in an area near 42°N-5°E where the density stratification is less stable than in the periphery. In the presence of important thermal losses suffered in winter by the sea surface, due to high winds (Mistral and Tramontane) made of dry cold air blowing over a warm sea (10°C more than the air at the coast), the density reached in the central area is greater than elsewhere and convective motions are likely to start there first. The buoyancy of the surface layer is very small and a "dense patch" appears, whose σ_θ is near that of the IW: it is the "preconditioning phase".

In such conditions (Medoc Group, 1970) high winds lasting for ten days and resulting in an evaporation of about 2 cm/day are able to homogenize the water down to 2000 m, during a "violent mixing phase". The vertical profiles of θ and S can be accounted for by vertical mixing. The mass of the interested water ($\sigma_\theta \cong 29.10$ from the surface down to more than 1000 m), occupied a volume, in February 1969, which intersected the surface over a more or less circular area centered at 42°N-5°E, and of about 60 miles in extent: the "dense patch"

(Tchernia, Fieux, 1971; Fig. 2*a*, *b*, *c*, *d*). However although the phase is a "violent mixing" one, a progressivity appears in the mixing during the interval between the two periods: February 10-12, 1969 and 17-20, while the "Discovery" was investigating in the "dense patch". This appears by comparing the characters of the superficial layers *homogeneous in salinity* ($S_m \pm 0.005^0/_{00}$) and thicker than 1 000 m (Fig. 3) during the two periods. The homogeneity reached its greatest depth (2 200 m) during the second period, February 17-20 (see Sankey, 1973). But, then, there is a decrease of the θ and S of the layer: while they were about 12.84°C-38.437$^0/_{00}$ during February 10-12, they shifted to 12.80°C-38.427$^0/_{00}$ during February 17-20; as for the density, it practically remained unchanged ($\sigma_\theta \cong 29.105$). As the latter water may be the result, as seen on a θ/S diagram, of mixing 2/3 of the former water with 1/3 of the "old" deep water below, one can conclude that a *continuing mixing process* took place in the interval, without any appreciable effect of surface phenomena, despite the occurring of gale-force winds (40 knots) for about two days, February 14-16 (Medoc Group, 1970).

So, even during the "violent mixing phase" a progressivity appears in the evolution of the "new" deep water, on a time-scale of a week.

Afterwards, the "spreading and sinking phase" takes place, with the partition of the "batch" of "new" deep water into fragments.

Mechanisms likely to be involved

Now, the short description of the phenomena given here (for more details, see Sankey, 1973) leaves open as well as does Sankey's study, the dynamics of the system. Killworth (1976) using two dimensional models

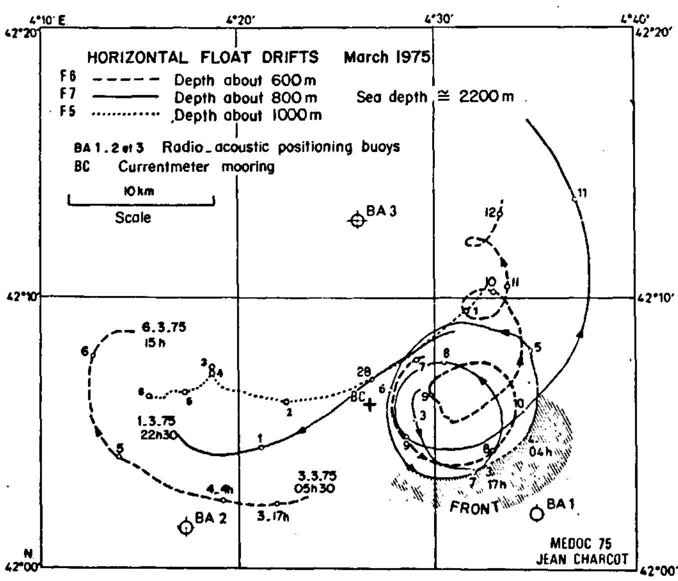

FIG. 4

Figure 4

Medoc 1975 cruise. Horizontal trajectories of deep floats (600, 800, 1 000 m) over depths of about 2 200 m. Acoustic positioning March 1-12, 1975 (Gascard, 1977).

Campagne Medoc 1975. Trajectoires horizontales de flotteurs profonds (600, 800, 1 000 m) au-dessus de fonds de 2 200 m environ. Localisation acoustique 1-12 mars 1975 (Gascard, 1977).

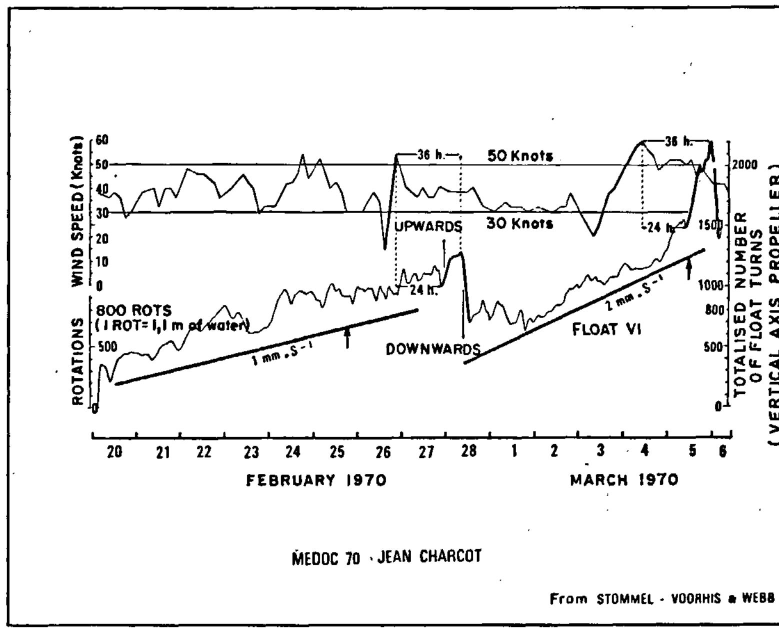

Figure 5

Vertical displacements recorded close to 42°N-4°20'E, at 400 m depth, as a function of time. Relation with wind velocities observed at Bouée-Laboratoire 1970 (Stommel, Voorhis, Webb, 1972).

Déplacements verticaux observés près de 42°N-4°20'E, à 400 m de profondeur, en fonction du temps. Relation avec le vent mesuré à la Bouée-Laboratoire 1970 (Stommel, Voorhis, Webb, 1972).

explained the *violent mixing* phase by a non-penetrative convection model and the *spreading phase* by a baroclinic instability process.

However, by tracking, Webb, Dorson and Voorhis (1970) modified VCM's floats 600, 800, 1 000 m deep, Gascard (1977; 1978) discovered and studied, for about one week, a deep eddy about 10 km in diameter which was likely to occupy most of the water column, a revolution lasted 2.5 to 3 days; the tangential speed was about 10-15 cm/sec. (Fig. 4).

By carefully analyzing the data from all the cruises 1969 to 1975 (except 1971), in the area (hydrology, eulerian, lagrangian and vertical current measurements) Gascard showed that, *at the end of the preconditioning phase*, a baroclinic instability of the flow on the Southern side of the dense superficial patch could generate meanders then eddies. The convergences and divergences associated with the growing meanders could cause intermediate water to move upwards, while "preconditioned" water of the central dense area of the cyclonic path could move downwards: mixing over a great range of depths then takes place, to about 1 200-1 500 m: baroclinic instability is one mechanism for building up thick homogeneous layers. He suggested that the eddies ($\emptyset = 10$ km) around the dense patch ($\emptyset \cong 100$ km $\cong 1°$ of latitude) were the counterpart at a scale of about 1/100 of atmospheric depressions (1 000 km) around the atmospheric polar front (100° wide $\cong$ 10 000 km).

However, as judged from vertical velocity records, other dynamical features are present. Studying a number of such records from 1970 to 1975 (Voorhis, Webb, 1970; Stommel, Voorhis, Webb, 1972) (Fig. 5), Gascard (1978) suggested mechanisms to explain the three ranges of vertical velocities recorded in the area:

(*a*) an aperiodic slow upward motion of 1 to 2 mm/sec.;

(*b*) periodic motions on several hours period in the range of 1-2 cm/sec.;

(*c*) intense short (2-3 hours) pulses of very high upwards and downwards velocities (up to 9-10 cm/sec.).

(*a*) The first motion is related to the slow ascent of intermediate water during the baroclinic instability process. It is not clear to what extent this measured slow ascent can fit Seung's (1979) results, which display, in addition to the winter time mean slow ascent (Fig. 1), a nearly three week period vertical motion of about 2 mm/sec. amplitude, modulating the mean ascent. The oscillation may be associated to the crossing, through the area, of eddies similar to those observed by Gascard;

(*b*) Gascard (1973) showed that the periodic motions were associated with stability oscillations on the local Brunt-Väisälä frequency, probably induced by bursts of wind.

(*c*) the third "catastrophic" vertical velocities which may appear one or two days after the onset of strong winds may be related to high internal waves generated by the wind drag or the wind surface cooling acting on a quasi-homogeneous ocean, according to Saint-Guily's model (1972; 1974).

Such high vertical motions were also measured in the Labrador Sea (Clarke, Gascard, 1979).

The mechanisms related to the two categories (*a*) and (*c*) of vertical motion, i. e. the baroclinic instability process and the large internal waves breaking, may both generate thick homogeneous layers which should have slightly different θ and S. Their mutual mixing, then their mixing with the "old" deep water contribute to small interannual variations of the Western Mediterranean deep water (see next chapter).

The motion (*a*) seems to be met under relatively good weather (observation of the eddies in 1975), the motions (*c*) under wind gales, 1970.

It is particularly disappointing that, due probably to battery failure, none of the 4 VCM floats which were involved in the tracking of the 1975 eddies (Fig. 4) obeyed the "order" of coming up to the surface, as they had satisfactorily done after the preceding "orders"... Thus, we had no access to their internal records.

Additional measurements and research are needed to completely clear up the various mechanisms which are at work in deep water formation.

THE INTERANNUAL COMPONENT

The interannual evolution of the characters of the "new" deep water (that of the winter) is not very easy to determine, for the range of variations is very small. In addition, as indicated above, the characters of the winter homogeneous layers change within a few days in a definite winter.

However, if the same kind of plot as in Figure 3 is made for the years in which Medoc cruises were carried out, it is possible to detect some interannual variations.

For instance, in 1970, with the same convention as for 1969, the superficial layers homogeneous in salinity

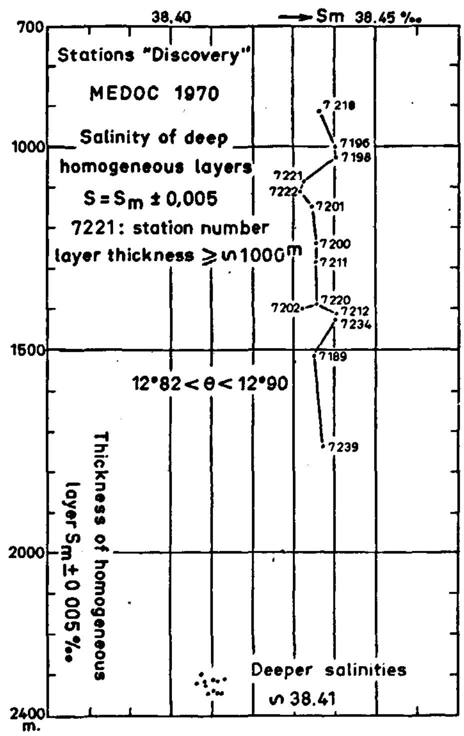

Figure 6

Salinities within layers homogeneous in salinity ($\pm 0.005^0/_{00}$) *as a function of their thickness (at least equal to* 1 000 *m from the surface). "Discovery" stations, Medoc* 1970.

Salinités au sein des couches homogènes en salinité ($\pm 0{,}005^0/_{00}$) en fonction de leur épaisseur (au moins égale à 1 000 m à partir de la surface), Stations « Discovery », Medoc 1970.

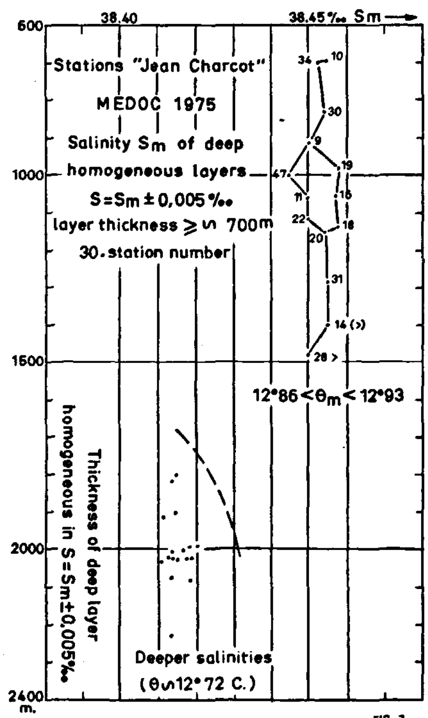

FIG. 7

Figure 7

Salinities within layers homogeneous in salinity ($\pm 0.005^0/_{00}$) *as a function of their thickness (at least equal to* 700 *m from the surface). "Jean Charcot" stations, Medoc* 1975.

Salinités au sein des couches homogènes en salinité ($\pm 0{,}005^0/_{00}$) en fonction de leur épaisseur (au moins égale à 700 m à partir de la surface), stations « Jean Charcot », Medoc 1975.

(within $\pm 0.005^0/_{00}$) and thicker than 1000 m have θ and S within the range 12.82°C-12.90°C, 38.43-38.44$^0/_{00}$, 29.10 in σ_θ (almost identical to those of the 1969 first period (February 10-12, 1969), while the deeper values are 12.71°C, 38.41$^0/_{00}$, 29.11 in σ_θ (Fig. 6);

In 1975 (Fig. 7), the θ and S of the homogeneous layers of thickness greater than 700 m were in the range of θ: 12.86-12.91°C; S: 38.45 to 38.46$^0/_{00}$. There is a distinctly higher θ and S than in the above-quoted years; the σ_θ's are near 29.105.

As for the deeper characters, θ is near 12.72°C, while S is near 38.415$^0/_{00}$ ($\sigma_\theta = 29.11$).

It appears as if the milder the winter, the greater the salinity and temperature in the mixed layers, but σ_θ varies very little.

We may recall here that during the very cold 1963 winter (January-March) (Lacombe, Tchernia, 1972 *b*), the deep homogeneous water was about θ = 12.80°C, S = 38.415$^0/_{00}$, $\sigma_\theta = 29.097$ (salinity by Knudsen's method), while the deeper water was 12.70°C, 38.405$^0/_{00}$, 29.11 in σ_θ.

In 1961, the deeper water was around 12.72°C, 38.400$^0/_{00}$ (Miller *et al.*, 1970).

A θ-S diagram (Fig. 8), although established on a slightly different basis, summarizes the results. It shows, for different years when measurements were taken, the evolution of the thick homogeneous layers, either superficial or *not*, and of the deeper layers ("old" deep water). For the curves 1973, see Figure 9. Results for 1974 appear on Figure 8. We did not mention this year earlier, because, at the late period of the cruise (after March 15, 1974), the homogeneous layer was no longer *superficial*.

Thus, the new deep water characters, as defined by the θ and S of the deep homogeneous (in salinity) layers, may change appreciably from year to year; the "old" deep water itself also changes, although on a smaller range from one year to the other.

In the other areas of deep water formation in the Mediterranean, some evolution of the deep water present has been evidenced by Mrs Zore-Armanda (1963) for the Adriatic. As for the "Levantine water", important variations occur, the greater as the thickness of the concerned layer is smaller (250 m maximum). For instance, in 1948, the "Atlantis I" (Conseil Permanent International pour l'Exploration de la Mer, 1952) found values of the Levantine Intermediate Water (LIW) of 15.8°C, 39.14$^0/_{00}$, while the same ship, in the same area in 1962, found 16°C, 38.98$^0/_{00}$ (Miller *et al.*, 1970).

The same is met in the Northernmost parts of the Aegean Sea where in the deeper layer (1500 m), in the summer 1955, we found: 12.7°C, 38.80$^0/_{00}$, 29.4 in σ_t (Lacombe, Tchernia, 1959) while, in 1948 (Conseil Permanent International pour l'Exploration de la Mer, 1952), in the same place, the "Atlantis I" had found 13.5°C, 38.79$^0/_{00}$.

The occasional bottom layer

Now, another phenomenon was found by P. Tchernia in 1972 and 1973, substantiated by J. Crease (pers. comm.).

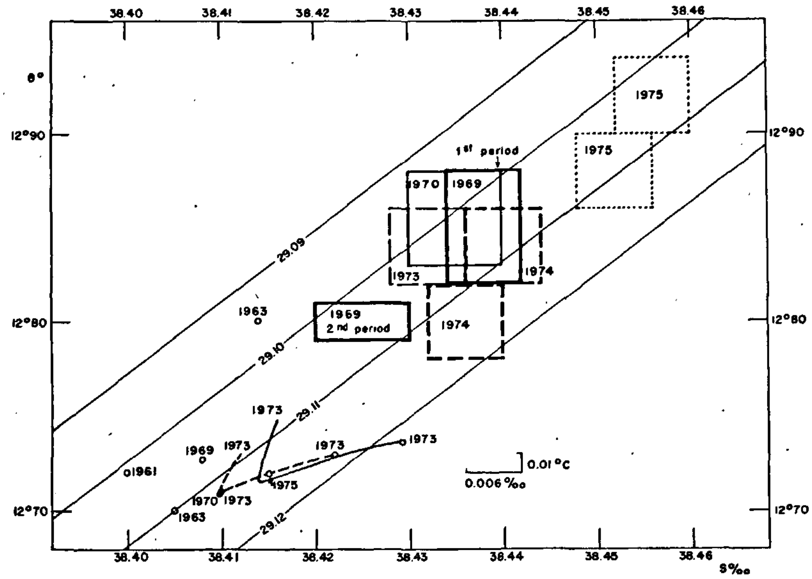

Figure 8

Ranges of characters of waters occupying the greater relative volumes in the Medoc area, from 1969 *to* 1975 *(except* 1971*). Circles indicate the deep "old" water characters for a number of years since* 1961. *The curves* 1973 *show the "occasional bottom layer".*

Gammes des caractères des eaux occupant les volumes relatifs les plus importants dans l'aire Medoc, de 1969 à 1975 (sauf 1971). Les cercles indiquent les caractères de l'eau profonde « ancienne » pour diverses années. Les courbes 1973 montrent la « couche de fond occasionnelle ».

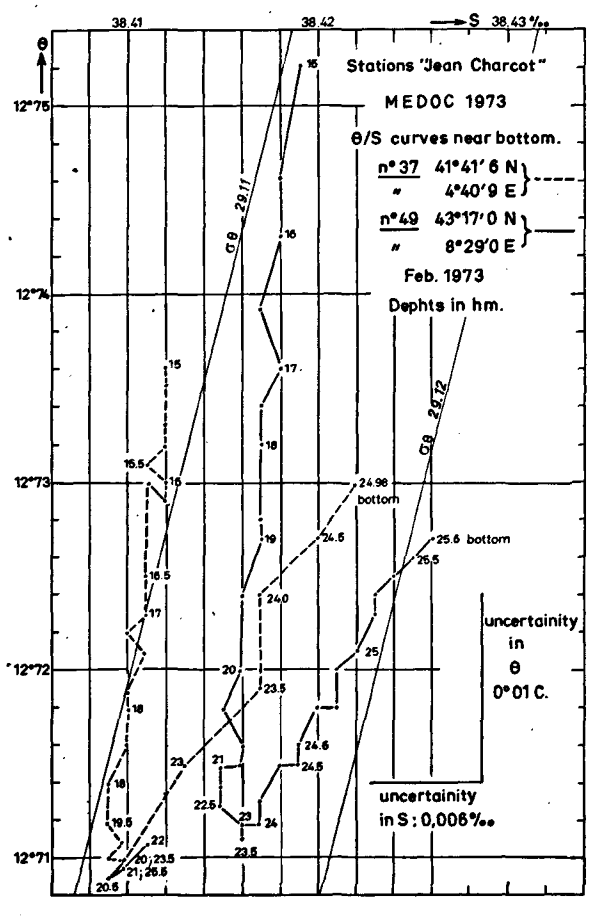

Figure 9

θ.S diagrams of the deeper layers in two cases of θ and S increase near the bottom ("Jean Charcot"/Medoc 1973).

Les diagrammes θ.S des couches profondes dans deux cas d'augmentation près du fond de θ et de S (« Jean Charcot »/Medoc 1973).

While, in the winters 1969, 1970, 1974 and 1975, in the Northern basin of the Mediterranean, both θ and S decreased very slowly, but continuously, between 2000 m and the bottom, in 1972 and 1973 (which were rather mild winters, although windy), there was a clear minimum of θ and S at about 2000 m; beyond, an increase of both θ and S occurred with positive gradients of the order of 0.005°C/100 m in θ and $0.005^0/_{00}$/100 m in S. In 1973, it was possible to check that the phenomenon extended over most of the Western basin North of the line: Balearic Islands-Corsica (see Fig. 9, and 8 lower left corner). The shift between the two curves is within the possible errors of absolute values and, at this stage at least, is given no significance. But the near-bottom increases have one. The situation in 1972 was comparable and the "Discovery" (J. Crease) found it in the Medoc area in November 1972; it should last one year at least. It was not present in February 1961 ("Atlantis I" cruise; Miller *et al.*, 1970) nor in 1963 ("Origny" cruise on a very cold winter; Lacombe, Tchernia, 1972 *b*).

The relatively high bottom salinity may be related to the characters of the new deep water which, as we have seen, is both slightly warmer and saltier at the end of rather mild windy winters (cases of 1972 and 1973). However, the high σ_0, reaching 29.12, is difficult to explain. It may be noticed that the level of the minimum of θ and S is higher in the West of the area concerned (2000 m) than in the East (2300 m). Whether these increases near the bottom can be related or not to the formation of very dense water on the shelf, such as was studied on the 1971 measurements by Person (1974), is not clear. The number of available data is at present too small for an adequate analysis of the phenomenon.

SOME FEATURES OF THE SPATIAL VARIABILITY

Spatial variability is, for a long-time scale, related to the distribution of the mean density field and to the geostrophic flow pattern. In fact, these factors are only badly known in most areas.

We shall only point out some cases of variability:

(*a*) The upwelling phenomena in the Gulf of Lions (Southern coast of France)

This area is subject to important upwelling phenomena under the effect of transient NW winds. But, contrary to the usual case in which the upwelling is driven by winds more or less parallel to the coast, the wind, here, is almost perpendicular, blowing seaward.

An interesting point (Millot, 1979), which first appeared from the examination of infra-red satellite thermographies, is that, after the onset of the wind, the upwelling first starts as point-sources along straight segments of the coast, about 10 km in length, while alongshore horizontal thermal gradients were located near salient points. Six source-points were localized along the coast. Current measurements have shown that cellular circulations around these points occurred. As time goes on, the size of the cold coastal areas increases, they overlap and may extend to the whole area over the continental shelf.

Crepon, Hua and Thomasset have initiated the development of a non-linear upwelling model, starting at the onset of a constant wind, in space and time, and blowing over a two-layer ocean. The necessity of taking into account the small-scale features of the coastline led them to use a finite element method for this numerical model. The preliminary results appear promising in showing the "point-source" generation (Hua, Thomasset, 1980).

(*b*) **As in the previous case,** infra-red imagery from satellites can help strongly in discovering meso-scale features in the Mediterranean area. For instance, the cold strip related to the anticyclonic Western Alboran Sea Gyre (Lanoix, 1974; Cheney, Doblar, 1979), is easily followed.

The Maltese front (Briscoe *et al.*, 1974) also gives very clear thermographies.

A very impressive survey of a number of meso-scale features, more or less regularly present in the Mediterranean, can be found in a recent paper (Philippe, 1980) from the "Centre de Météorologie Spatiale" of Lannion.

CONCLUSION

It may be stressed that the size of the Mediterranean and the subsequent possibility of monitoring it with relatively modest facilities (which, however were never implemented systematically, except in limited areas), tend to prove that this sea may be a good field for studying the ways in which the sea responds to atmospheric and energy exchange forcing.

From what we know, it is clear, that the mechanisms involved in deep water formation in the Medoc area are present in other open-sea areas of deep water formation, in particular in the Labrador Sea (Clarke, Gascard, 1979). The discovery by Gordon (1978), in the Weddell Sea, of a structure analogous to a 14 km-eddy system, seems to indicate that the same processes may also act in the Antarctic.

The fact, also recognized (Fieux, 1978), that the evolution of the sea-surface temperature in the Mediterranean, since the beginning of the century, is identical with that found for all areas of the N. hemisphere, which have been studied in this respect, demonstrates that this sea may be a "pilot" basin for that time scale. It may prove one for the climatological scale (Béthoux, 1978).

We have at our disposal a "model" basin in which many oceanic phenomena may be studied easier than elsewhere. We should devote a greater attention to it, in particular for studying meso-scale problems which present conspicuous features in the area.

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