Wind-induced upwellings and currents in the gulfs of Patras, Nafpaktos and Korinthos, Western Greece

Greek straits
Sea surface temperature
Upwellings
Eddies
Baroclinic motions

Détroits grecs
Température superficielle
Remontées d'eau
Tourbillon
Mouvement barocline

Alexander LASCARATOS [a], **Ettore SALUSTI** [b], **Georgia PAPAGEORGAKI** [a]
[a] University of Athens, Department of Applied Physics, Laboratory of Meteorology, 33 Ippocratous Str., 10680 Athens, Greece.
[b] Istituto Nazionale de Fisica Nucleare (INFN), Department of Physics, Universita "La Sapienza", P.A. Moro, I-00185, Roma, Italia.

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ABSTRACT

With the use of 152 NOAA-6 and NOAA-7 satellite infrared images taken over the gulfs of Patras, Nafpaktos and Korinthos in Western Greece, a relatively complete description of sea surface temperature features and their evolution in time is obtained for these three marine basins. The satellite data, discussed in the light of recent theories on wind-induced marine upwellings near capes (Crépon and Richez, 1982 and Crépon *et al.*, 1984), allow us to provide a general picture of wind-induced movements and upwellings.

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RÉSUMÉ

Remontées d'eau induites par le vent et courants dans les golfes de Patras, Nafpaktos et Corinthe, Grèce

La structure thermique superficielle des golfes de Patras, Nafpaktos et Corinthe (Grèce) et son évolution sous l'effet du vent sont présentées. Les données de 152 thermographies des satellites NOAA-6 et NOAA-7 sont confrontées aux théories récentes sur les remontées d'eau au voisinage des caps (Crépon et Richez, 1982; Crépon *et al.*, 1984); elles conduisent à une description générale des mouvements marins et des remontées d'eau dues aux vents dans la région.

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INTRODUCTION

In this paper, the sea surface temperature features observed in infrared satellite images of the gulfs of Patras, Nafpaktos and Korinthos (Western Greece) are discussed. The satellite data, comprising 152 infrared images taken from June 1981 to November 1982 by the NOAA-6 and NOAA-7 satellites, are examined in relation to wind data and are ultimately interpreted in the light of the models of Crépon and Richez (1982) and Crépon *et al.* (1984).

The sea water body studied lies between the Greek mainland and the Peloponnesus (Fig. 1) and comprises the two main gulfs of Patras and Korinthos as well as the small bay of Nafpaktos connecting them. The Gulf of Patras has an almost circular shape with maximum depths at the centre of 120 m and a diameter of 20-25 km. It merges in the west with the Ionian sea through an opening of 15 km wide and 50 m deep between Cape Pappas and the Messolongi lagoon. At the eastern end, the gulf runs into the Bay of Nafpaktos through the strait of Rio-Antirio, which has a width of 2 km and a sill depth of 60 m. This small bay is almost circular in shape, with a diameter of 7 km and a maximum depth of 100 m. East of the bay lies the strait separating Capes Mornos and Drepanon, which is 3.5 km wide and has a cell depth of 90 m, and constitutes the entrance of the Gulf of Korinthos. This large water body has an E-W extension of 130 km, a N-S extension which varies from 10 km in the western part to 30 km in the eastern, a maximum depth of 900 m at the centre and very steep coasts. Finally, in

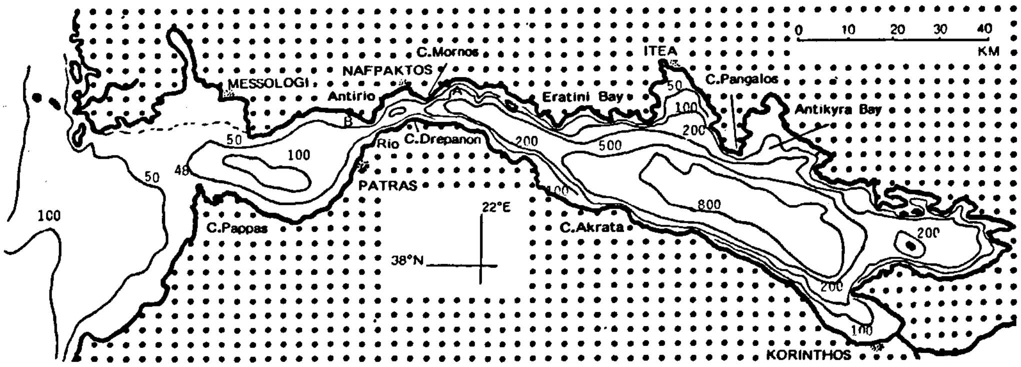


Figure 1
General map of the studied zone (depths in metres).

the extreme southeast, through a very narrow artificial channel – the Korinthos Canal – the Gulf of Korinthos runs into the Saronikos Gulf.

The hydrological properties of these two water bodies were examined by Friligos *et al.* (1985), for the Gulf of Patras and by Anderson and Carmack (1973) for the Gulf of Korinthos. It is interesting to note that the Gulf of Patras is an area where mixing processes take place (Friligos *et al.*, 1985). More saline water (38.6-38.7) enters from the Ionian Sea, while less saline water (38.3-38.5) comes from the Gulf of Korinthos. The dense water filling the central trough of the Gulf of Patras seems to be isolated. The vertical temperature gradient $dT/dz = q$ for the upper layer is 22×10^{-2} °C/m in summer. According to Anderson and Carmack (1973), the temperature of the Gulf of Korinthos varies: at the surface from 25°C in summer to 14.3°C in winter; below 200 m remaining approximately equal to 13.3°C in summer and 12.8°C in winter. The vertical temperature gradient q for the upper layer is therefore 12×10^{-2} °C/m and 10^{-2} °C/m in summer and winter respectively. The salinity of the waters below 200 m remains unchanged, 38.38 in winter and 38.42 in summer. Their T-S diagrams show that:

a) there is an entrance of Ionian waters above the Rio-Antirio sill towards the Gulf of Korinthos;

b) during summer the waters deeper than 250 m in the gulf are isolated. Winter convection nevertheless seems to reach the deepest parts of the gulf. Thus, although the gulf acts as a nutrient trap, the nitrate and phosphate concentrations are less important than those observed in anoxic basins.

Large tidal currents are observed in both the Rio-Antirio (maximum speed 100 cm/s) and Mornos-Drepanon straits (maximum speed 60 m/s, Hellenic Hydrographic Service, Pilot, 1984).

The area is often disturbed by Northwestern and Northeastern winds. The high mountain chain on the western Greek mainland, together with the high and abrupt mountains on both sides of the Rio-Antirio strait, further funnel these winds and change their direction inside and close to the strait to WSW and ENE, *i.e.* along the axis of the channel. During strong synoptic situations, the wind speed in the straits can easily reach 15 m/s with a mean value of 7 m/s. This gives an average curl of wind of $1.4 \times 10^{-3}\,s^{-1}$ for space scales of 5 km, which is reasonable considering the surrounding steep mountains. In Figure 2, we give a synopsis of the resultant marine currents and upwellings, important for fisheries and pollutant transports as well as interesting from the theoretical point of view.

Theoretical studies by Crépon and Richez (1982) and Crépon *et al.* (1984) relative to capes and coasts of open oceans may be found in the literature. Furthermore, there are many studies of wind-induced currents in marine straits, but the effect of wind-induced upwellings in marine straits, at least for this part of Mediterranean Sea, is a novel theme.

We first analyse the data; then discuss the main sea surface temperature observed in infrared images; and the comparison with the results of theoretical models.

In general a good agreement is found between our observations and the prediction of the theoretical models.

Data

For this investigation, 152 infrared NOAA-6 and NOAA-7 thermal images received at Lannion station (France) covering the period from June 1981 to November 1982 were used. Each photograph has 16 grey tones corresponding to relative temperature changes. The full scale of tones corresponds to 8°C: each tone differs, therefore, from its surroundings by 0.5°C. This relative scale of temperatures gives better accuracy than absolute scaling, at best 1 or 2°C (Beer, 1983). For each photograph the lines separating the consecutive tones were drawn, with numbers showing the relative cooling, namely the departure from maximum temperature.

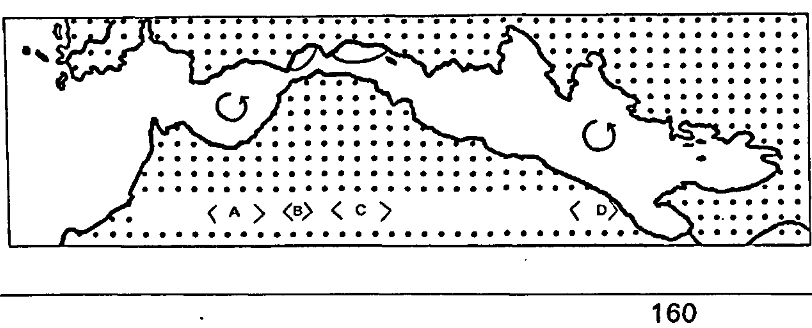
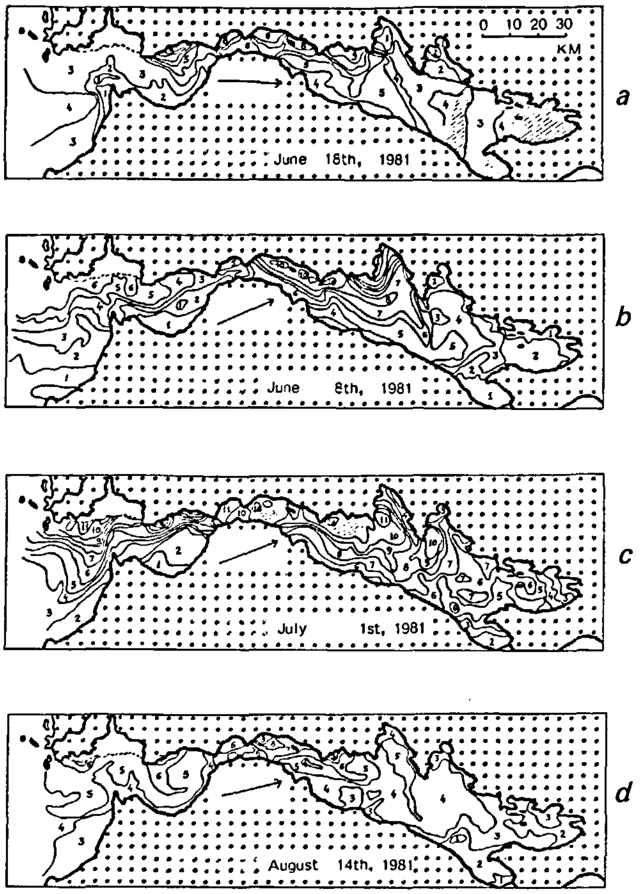


Figure 2
Schematic representation of the main thermal regularities of sea surface temperature, observed in thermal satellite images of the work zone: A: cyclonic eddy in the Gulf of Patras; B: upwelling in the Bay of Nafpaktos; C: upwelling in the Gulf of Korinthos; D: cyclonic eddy in the Gulf of Korinthos.



Twelve images were selected to present the most interesting sea surface temperature features observed (Fig. 3 *a-l*); some of these images will be referred to more than once in the next.

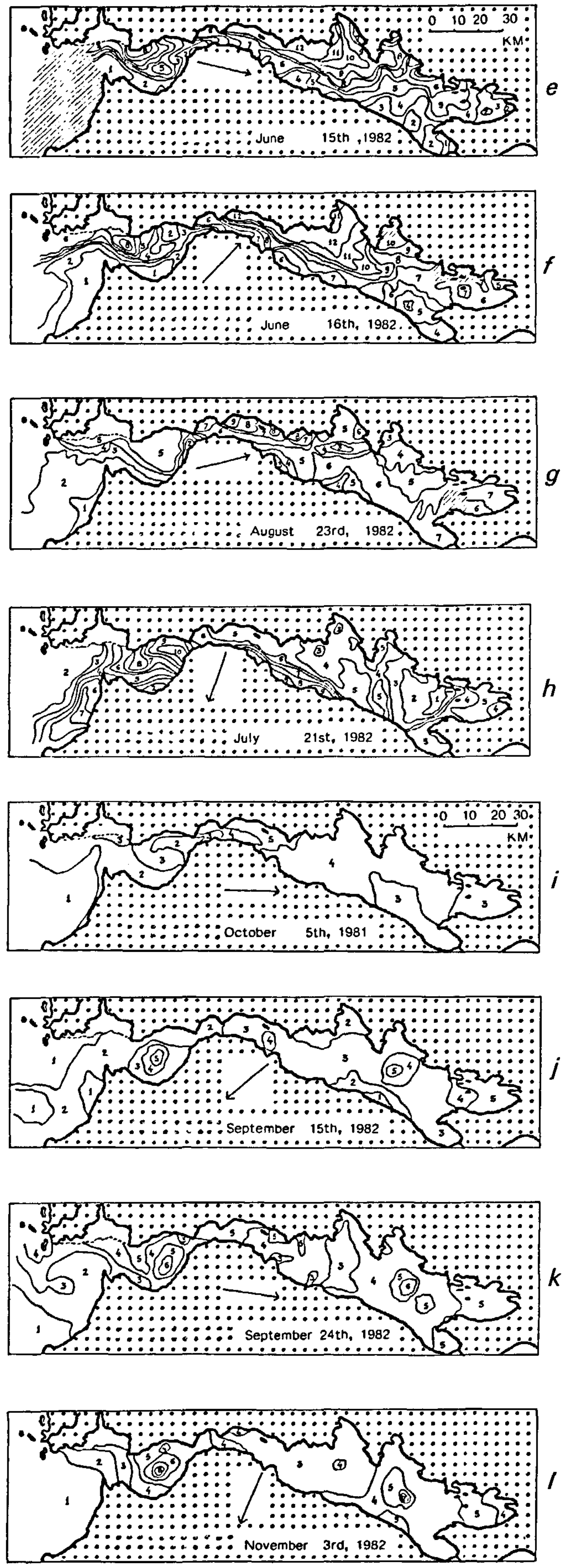
Wind data were obtained from National Meteorological Service (stations at Patras and Nafpaktos) and from a station of the University of Patras at Rio.

Main sea surface temperature features

Gulf of Patras

All year round the temperature of the surface waters of this gulf is equal to or lower than that of the open Ionian Sea. This difference is smaller in autumn and winter (Fig. 3 *d*, 3 *g*, 3 *i*, 3 *l*); at the begining of March a thermal front separating the Gulf of Patras from the Ionian sea is observed. In spring, summer and autumn a temperature gradient between the northern and southern coasts of the gulf exists due to the entrance in the southern banks of warmer Ionian Sea waters (Fig. 3 *b-g*). They enter the gulf north of Cape Pappas, following a cyclonic movement. There is no N-S gradient of temperature inside the gulf during winter.

A characteristic phenomenon observed during both of the summers studied is the occurrence of cold waters in the N-NW coasts of the gulf (9 out of 29 summer photographs during 1981) with the isotherms oriented parallel to the north coasts and 2.5-3°C temperature difference (Fig. 3 *b-c*). In summer 1982, this phenomenon was observed in 6 out of 37 photographs and the N-S temperature difference was 1.5 to 3.5°C (Fig. 3 *e*).



Figures 3 *a-l*
Twelve images representing infrared pictures. Each line separates two areas with a temperature difference of 0.5°C. Numbers show the cool departures from maximum temperature (number 1). Wind directions are also shown.

At the Gulf of Patras during late spring and summer, circular tongues of water are frequently observed (Fig. 3 *d*). The warmer waters originate from the Ionian. Entering through Cape Pappas, they follow the circular shape of the coast and move to the north-east. During the autumn, a cold patch at the centre of the gulf is permanently present. The isotherms have a clearly closed circular shape. The temperature difference between the centre and the circumference is 0.5-2.5°C (Fig. 3 *j*, *k*, *l*), showing that the thermocline has risen. In autumn 1981, the patch appears in 9 out of 14 images (Fig. 3 *i*), while in autumn 1982 it appears in 21 out of 22 images. In autumn 1982, this patch is particularly well defined. According to Friligos *et al.* (1985), in August 1980 the surface temperature of the Gulf of Patras varied from 24.5 to 25.8°C, but in its central part the temperature was 23.9°C. When cold waters appear in the northern coasts of the gulf, the cold patch tends to disappear.

Finally, during both summers, cold waters can sometimes simultaneously be observed in the vicinity of cape Pappas (west of the gulf entrance) and near Patras city. The longest of such events occurred from 20 to 22 July 1982 and is related to specific sea surface temperature structures in the Gulf of Korinthos (Fig. 3 *h*).

Nafpaktos Bay

The small Bay of Nafpaktos connects the two major gulfs of Patras and Korinthos. During autumn, winter and spring, the bay is approximately 0.5-1°C cooler than both surrounding gulfs. In summer, because of the presence of important thermal fronts in the area, no such clear relationship can be stated.

Inside the bay during summer, a sea surface temperature difference is often observed, with colder waters to the north in the immediate vicinity of Antirio. The appearance of these cold waters inside the bay coincides with the appearance of cold waters in the north coasts of the two neighbouring gulfs. Inside the bay, the N-S sea surface temperature difference reaches 1.5-2°C (Fig. 3 *a*, *b*, *e*, *f*).

Gulf of Korinthos

There is an almost permanent E-W sea surface temperature difference inside the gulf, the warmer waters being in the eastern part. In summer, this difference reaches 4°C and in autumn 1.5°C. In winter it almost disappears, while in spring the difference appears again. The waters in the bays of Itea and Anticyra are almost permanently warmer than the rest of the gulf.

A particularly stimulating phenomenon consists in the appearance of very cold waters on the NW coasts of the gulf, namely from Cape Mornos to Eratini bay (16 out of 28 photographs in summer 1981 and 14 out of 36 photographs in summer 1982). This thermal front has isotherms almost parallel to the coasts (Fig. 3 *a*-*g*). The sea surface temperature differences are quite remarkable: 1.5-7°C in summer 1981 and 2-6°C in summer 1982. In some cases, the appearance of cold waters in the NW part of the gulf is accompanied by a simultaneous appearance of cold waters on the SW edges of the Itea and Antikyra bays (Fig. 3 *c*).

During both summers, an inverse phenomenon, *i.e.*, the occurrence of colder waters on the southern coasts of the gulf (more specifically from Cape Drepanon to Cape Akrata), is sometimes observed. These events are related to similar sea surface temperature structures in the Gulf of Patras (Fig. 3 *h*).

Finally, during autumn 1982, a cold patch is frequently observed in the middle of the gulf, south of Antikyra bay. The sea surface temperature difference between the colder centre of the patch and its borders is 1-1.5°C, for a distance of 7 km (Fig. 3 *j*-3 *l*). This phenomenon coincides in time with the occurrence of the cold patch in the centre of the Gulf of Patras.

Theoretical discussion

The cyclonic eddy in the Gulf of Patras

From satellite images, a permanent thermocline surfacing of the waters inside the gulf, from spring through autumn, is evident. Assuming that the upper layer is more energetic than the lower, the resulting motion is cyclonic (Defant, 1961). In winter the small difference of the sea surface temperature does not permit any conclusions to be drawn.

On theoretical grounds, one could expect that the interaction of the wind curl with bottom topography gives rise to a barotropic cyclonic motion, coupled with a baroclinic motion. A classical discussion for circular flows is found in Defant (1961; *see* also the synthesis of Neumann and Pierson, 1966), showing a thermocline surfacing similar to the one observed in this gulf.

Let us estimate the horizontal velocity of the upper layer. The data provides the following information:

– $\Delta T = 2.5$°C (horizontal surface temperature difference);

– $r = 7$ km (eddy radius);

– q in summer $= dT/dz = 22 \times 10^{-2}$ °C/m (vertical temperature gradient).

Assumong a simple two-layer picture, the upper layer rising makes an angle with the horizontal:

$tg\,\theta = (1/q) \times (\Delta T/r) = 1.6 \times 10^{-3}$.

For this two layer approximation of the Gulf of Patras, a $\Delta\rho/\rho$ value of 10^{-3} is taken, based on the results of Friligos *et al.* (1985). Assuming $g = 10\,\mathrm{m/s^2}$ for the gravity acceleration and $f = 10^{-4}\,\mathrm{s^{-1}}$ for the Coriolis parameter, the upper layer velocity u_u and the lower layer velocity u_l must obey (Neumann, Pierson, 1966):

$$tg\,\theta = -\frac{f}{g}\,\frac{\rho\,(u_l - u_u)}{\Delta\rho}\left(1 + \frac{u_l + u_u}{fr}\right).$$

Taking first that $u_l = 0$, we have $u_u = 14$ cm/s. If, as a realistic estimate, we put $u_l = 5$ cm/s or $u_l = 10$ cm/s, one respectively obtains $u_u = 18$ cm/s or $u_u = 21$ cm/s, which constitute reasonable values for the region.

Wind induced upwellings

As already mentioned, in summer 1981 and 1982 (but also in some spring photographs), cold waters can be observed on the northern coasts of the gulfs of Patras, Korinthos and Nafpaktos. An examination of the winds in the Gulf of Patras shows a clear relation between the wind direction and these events, appearing one or more days after strong winds blowing from WSW. For the Gulf of Korinthos, a similar relation seems to exist: the winds in Rio (summer 1981) or in Nafpaktos (summer 1982) had a WSW direction a few days before each event. This supports the hypothesis that the observed phenomenon is essentially due to a wind-induced upwelling. As far as the occurrence of cold waters on the southern coasts of these gulfs is concerned, these events seem to be related to NE winds. Especially from 20 to 22 July 1982, strong winds were blowing from the North-East, a direction suitable for an upwelling to occur in the southern coasts, as shown in Figure 3 *h*.

The structure of these cold patches will now be compared with the results of the Crépon and Richez (1982) and Crépon *et al.* (1984) models (*see* Appendix). Their theoretical considerations are mainly based on the propagation of coastal Kelvin waves. Let us consider the line connecting point A and point B on the northern coasts of the gulfs of Patras, Nafpaktos and Korinthos (Fig. 1). This line is a combination of two consecutive capes of triangular shape, namely Cape Antirio and Cape Mornos.

Taking $h_1 = 30$ m as the upper-layer thickness and $h_2 = 40$ m as the lower-layer thickness in the coastal zone, it is deduced from Lascaratos *et al.* (1987) that $\Delta\rho/\rho = 10^{-3}$ is a good approximation. Therefore the Rossby internal radius of deformation results to be $r = 4$ km, while the speed of the baroclinic mode is $c = 0.4$ m/s (*see* Appendix). The overall length of the coasts from point A to B is 32 km: a Kelvin wave requires = 20 hours to progress from one point to the other. The sea surface temperature photographs are one per day and rarely two per day. Consequently the westward movement of the Kelvin waves and the alternating increase or decrease of the upwellings cannot be examined. Furthermore, the time needed for the creation of the K 2 Kelvin wave (*see* Appendix) is $t^* = 1/f \approx 3$ hours: we cannot expect to detect it through the satellite images. We can thus only examine whether the conclusions of Crépon *et al.* (1984) regarding triangular capes are confirmed by our sea surface temperature observations. As long as the entrainment of the cold lower-layer water into the surface layer can be disregarded, Crépon *et al.* (1984) show that after Kelvin waves have moved away, a constant difference Δh between the downwind and the upwind interface elevations persists, with upwelling intensity greater on the downwind than the upwind coast (*see* Fig. 10 and Fig. 13, of Crépon *et al.*, 1984, *see* also Appendix). In our case for a $0.1\,\mathrm{N/m^2}$ wind stress Δh is computed to be 7 m.

Our Figure 3 *a*, which represents an early stage of an upwelling event, shows small patches of colder water which are first observed downwind, in the immediate vicinity of the capes. As the phenomenon progresses, the isotherms stretch out in a direction more or less parallel to the downwind coast. The upwelling signal is stronger along the downwind coasts: the waters east of Cape Mornos are colder than those of the Bay of Nafpaktos which, in turn, are colder than those west of Antirio (Fig. 3 *b-g*).

A similar effect can also be observed when upwellings occur along the southern coasts. In fact, during the 20-22 July 1982 event, we see (Fig. 3 *h*) that the upwelling is again stronger along the downwind coasts of the triangular capes Pappas and Rio (note that the wind is now from NE).

More quantitatively, thermistor-chain measurements (Lascaratos *et al.*, 1987) immediately downwind from Cape Mornos, show an interface rise of 5-10 m, one day after the onset of the wind, and of 15 m approximately two days after, for winds of 7-10 m/s, blowing from WSW. They compute vertical velocities of 0.1 mm/s. On the other hand, from Crépon *et al.* (1982) theory, for parameters similar to our case we obtain vertical velocities of 0.14 mm/s (*see* Appendix, gulf of type B). As may be seen in Figure 3 *a-g*, the colder waters first appear in a zone adjacent to the coast, with a width of 5 km. This is also in agreement with Allen's continuously stratified model (1973) since the Rossby radius of deformation is $r \simeq 4$ km.

It may therefore be concluded that our observations are in good agreement with the theoretical predictions of both Crépon and Richez (1982) and Crépon *et al.* (1984), as far as the end stage of the upwellings is concerned.

The cyclonic eddy in the Gulf of Korinthos

This steady eddy in the Gulf of Korinthos observed during autumn 1982 (Fig. 3 *j-l*) is of unknown origin. It remains in the same area in the centre of the gulf, south of Antikyra Bay, where the depth exceeds 800 m. The wind data available in the area do not show any special characteristics during the days the eddy is observed. Making again the logical assumption that the upper layer's movement is more energetic, we conclude that the movement is cyclonic. The eddy's radius is ~ 7 km, of the same order as the baroclinic Rossby radius of deformation.

Since it is observed in the centre of the gulf, above its deepest part without important wind systems pre vailing when it occurs, the eddy is probably simply the surface signature of a permanent cyclinic circulation inside the gulf. Making the same calculations as in paragraph 4 *a*, for $\Delta T = 1°C$, $r = 7$ km and q in summer $= 12 \times 10^{-2}$ °C/m, the slope of the interface is 1.2×10^{-3}. Assuming that the lower layer is at rest, one has an estimate of the surface horizontal velocity $u_u = 11$ cm/s.

SUMMARY

The sea surface temperature features present in the gulfs of Patras, Nafpaktos and Korinthos have been studied using 152 infrared satellite images covering a period of 18 months. The most interesting and frequent features observed include (Fig. 2).

a) the existence of cyclonic eddies in both gulfs, mainly during autumn. The eddy of the Gulf of Patras seems to be enhanced by barotropic/baroclinic transports;

b) the occurrence of a strong and well-defined upwelling on the northern coasts of the area during the summer period. The structures of the observed upwellings are in good agreement with the theoretical predictions of Crépon and Richez (1982) and Crépon *et al.* (1984).

Acknowledgements

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REFERENCES

Allen J. S. (1973). Upwelling and coastal jet in a continuously stratified ocean. *J. phys. Oceanogr.*, **3**, 245-257.

Anderson J. J. and E. C. Carmack (1973). Some physical and chemical properties of the Gulf of Corinth. *Estuar. coast. mar. Sci.*, **1**, 195-202.

Beer T. (1983). *Environmental Oceanography*, Pergamon Press.

Crépon M. and C. Richez (1982). Transient upwelling generated by two-dimensional atmospheric forcing and variability in the coastline. *J. Phys. Oceanogr.*, **12**, 1437-1457.

Crépon M., C. Richez and M. Chartier (1984). Effects of coastline geometry on upwellings. *J. Phys. Oceanogr.*, **14**, 1365-1382.

Defant A. (1961). *Physical Oceanography*, Pergamon Press.

Friligos N., A. Theocharis and D. Georgopoulos (1985). Preliminary chemical and physical observations during summer 1980 on a silled embayment in the Ionian sea. *Vie Milieu*, **35**, 2, 115-125.

Hellenic Hydrographic Service, Pilot 1984.

Lascaratos A., E. Salusti and K. Nittis (1987). Packets of internal waves in the Gulf of Korinthos, Greece (in preparation).

Neumann G. and W. Pierson (1966). *Principles of Physical Oceanography*. Prentice-Hall, Inc.

Appendix

We briefly discuss below one of the basic conclusions of the work of Crépon and Richez (1982): "Transient upwellings generated by two-dimensional atmospheric forcing and variability in the coastline".

In the case of a gulf of type B (Fig. A 1) an upwelling and a coastal jet in the parallel to the wind coast are generated. The slope of the interface along the coast parallel to the wind is deflected on the other coast, and is then travelling to the south as a Kelvin wave. The speed of the baroclinic mode is

$$c^2 = g\frac{\Delta\rho}{\rho}\frac{h_1 h_2}{h_1+h_2},$$

while the Rossby internal radius of deformation is $r=c/f$. The upwelling rate is

$$h_t = \frac{\tau_y}{\rho c}\frac{h_2}{h_1+h_2}.$$

In a following work, namely Crépon *et al.* (1984): "Effects of coastline geometry on upwellings", several geometries of capes as well as combinations of all previous geometries are studied.

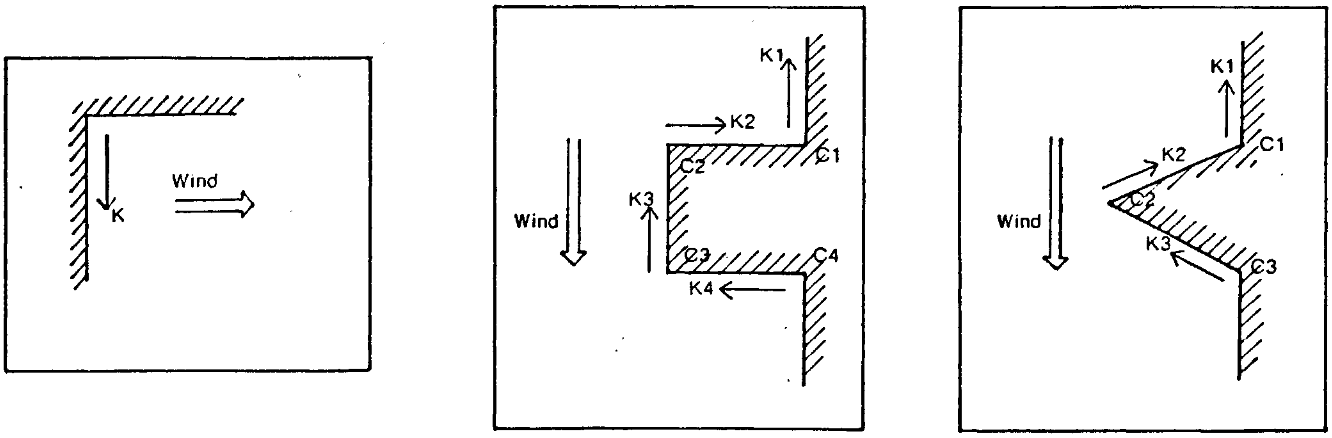
For combination of capes and gulfs of types A and B (Fig. A 2), Kelvin waves are formed in all corners. Thus Kelvin wave K 4, after it reaches corner C 3, travels to the north until it reaches C 2, then travels to east up to corner C 1, and then travels again to the north. All subsequent Kelvin waves will travel in the same way. After they have all passed, there will remain a difference

$$\Delta h = \frac{\tau_y}{\rho c^2} 2L \frac{h_2}{h_1+h_2}$$

in interface elevations of coasts C 1 − C 2 and C 3 − C 4. This difference is not due to the early downwelling which occurred along the coast C 1 − C 2, but is the result of propagation along the northern coast. It remains constant with time and uniform between any point of coast C 1 − C 2 and any point of coast C 3 − C 4.

For a triangular cape (Fig. A 3), two distinct periods of time can be distinguished: a) $0<ft<1$: during this short period, there is a response to the normal to the coastal component of the wind. Kelvin waves K 1 and K 3 are generated. A small downwelling appears in coast C 1 − C 2 while upwellings appear on the other coasts. b) $ft>1$: now the response is to the parallel to the coast component of the wind. The downwelling disappears and an upwelling takes its place. The Kelvin wave K 2 is generated. The signature of the Kelvin front passage is a difference of interface elevation between the up and the downwind coasts.

The main feature, stressed in the previous discussion, is that the strength of the upwelling is more important on the downwind than on the upwind coast.



Figures A. 1-A. 3
Three images showing the direction of the Kelvin waves' propagation, according to Crépon and Richez (1982) *and Crépon* et al. (1984).