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# ASYMMETRY AND ANOMALIES OF CIRCULATION AND VERTICAL MIXING IN THE BRANCHING OF A LAGOON-ESTUARY

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## ABSTRACT

The lagoon Ebrié, in Ivory Coast, is formed of different bays and branches before communicating by an artificial canal with the Gulf of Guinea. The two principal lateral branches are quite different : the western channel forms a natural prolongation of the central channel while the eastern channel begins with a constriction. Observed at the eastern and western entrances of the lagoon, the circulation is statistically different, particularly in the upper layer and during ebb-tide. The eastern channel shows, at times, anomalies of residual velocity profiles which determine the relative asymmetry : there is often a seaward jet in the mid layer. The stronger residual anomalies are connected with sensible departures from the semi-diurnal period, involving the existence of beats between the tides and other subtidal frequencies. A decrease of Richardson number occurs during the anomalous profiles. The vertical mixing, its asymmetry and anomalies could be explained by a criterion for the maintenance of turbulence, depending on transient stages of river discharges and on the wind at the subtidal frequencies.

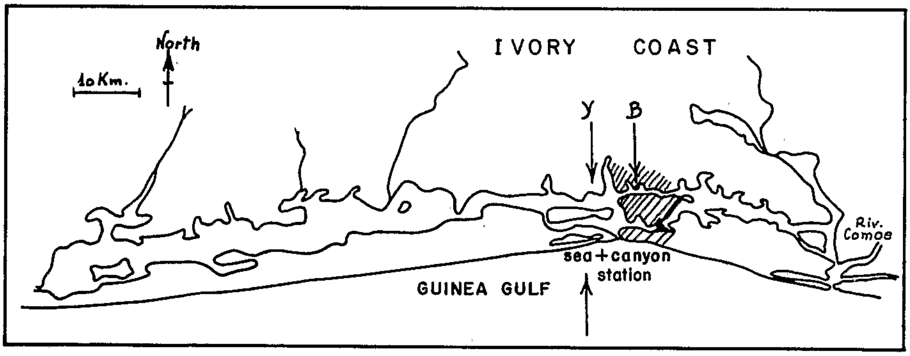



Fig. 1 - LAGOON EBRIE -Locations of temperatures - salinities - currents observations.

INTRODUCTION

The study of horizontal transport and of vertical mixing at the principal eastern and western gates of the lagoon Ebrié (fig. 1) was included in a larger and previously begun environmental program at the "Centre de Recherches Océanographiques d'Abidjan" in Ivory Coast.

Geophysical and physical descriptions (TASTET, 1974) of the whole lagoon have indicated its morphology, the monthly fresh water inflows, the tides and their currents. Because of the complicated topography with its sills, constrictions and bays and because of the variable fresh water inflow, the tides in the lagoon are much damped down and often out of phase with respect to the oceanic tides. TASTET (1974, p. 18) observes that these phase differences may reach respectively 40 minutes and 2.6 hours in the Central Channel, not far from the entrance of the eastern Channel. The seaward flow measured in the canal, during a tidal cycle of june 1966 was more than 400 $m^3 s^{-1}$, value which corresponds well to the mean fresh water inflow of June 1970, 1971, 1972 (TASTET, 1974, tab. 2, 3, 4).

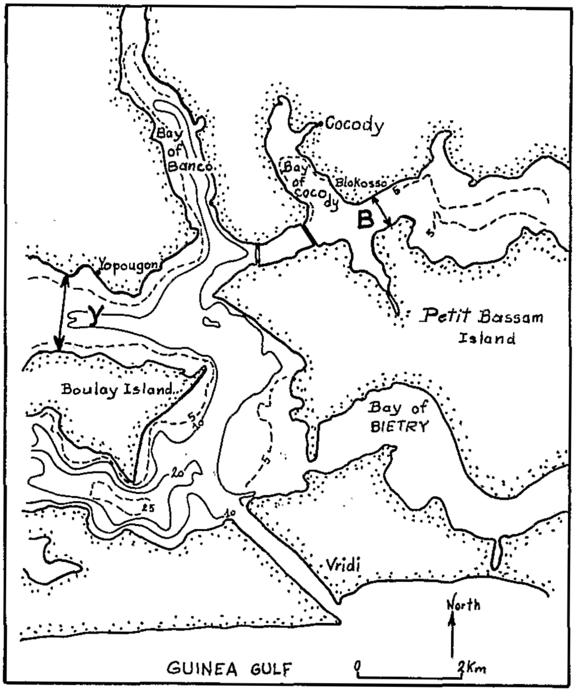



Fig. 2 - Geography and isobaths of the inner estuary
(from J.P. TASTET, 1973)

From January to July 1976, current, salinity and temperature measurements were made, every month, at the locations B and Y of fig. 2 : until April, EKMAN current-meters, sampling five minutes, every two hours, were completed with a mooring of two AANDERAA currentmeters recording integrated velocity, direction, temperature and depth every ten minutes.

From August to December, the program was modified and concentrated in the eastern channel, in order to obtain some tangible results between the circulation, averaged over several tidal cycles, and the estimations of fresh-water inflow.

EAST-WEST ASYMMETRY

A glance at the fig. 2 indicates the complexity of the estuarine morphology ; however, the net flow coming from the canal must reach the area of branching without sensible loss because of the large and deep central channel. Since both branches represent approximately equivalent areas, and under the assumption that the water surface remains horizontal, it could be supposed that the velocities $V_y$ and $V_b$ would be inversely proportional to the surface of the vertical sections $S_y$ and $S_b$ ; that would lead to the following relationship:

$$V_y = (S_b/S_y)\ V_b = 0.53\ V_b$$

$V_y$, $S_y$, $V_b$, $S_b$, are the tidal velocities and the vertical sections at the locations y and b.

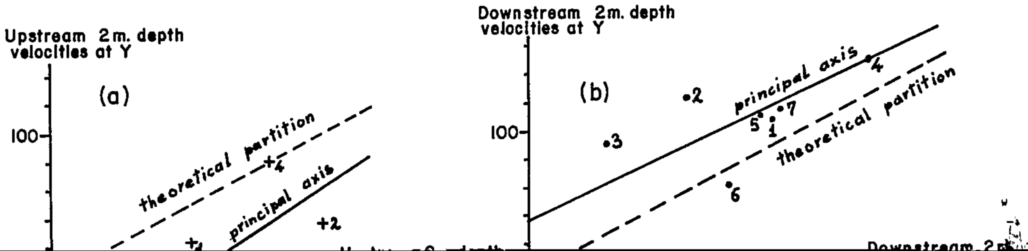
The observed velocities, summed over ebb or flow periods, are distributed along a principal axis expressed by

$$\frac{y - \overline{y}}{s_y} = \frac{x - \overline{x}}{s_x}$$

where $\overline{x}$, $\overline{y}$ are the means, $s_x^2$ , $s_y^2$ the variances of the summed velocities x and y.

Fig. 3a, b, c, d, show our results at the depths 2 m and 4 m, for ebb (downstream velocities) and flow (upstream velocities) compared to the simple model described above ; the upper layer deviates eastwards during flowing tide (fig. 3 a) ; during ebb-tide there is an excess of velocities in the upper layer and a loss in the mid layer (fig. 3 b, d) of the western section ; the mid flow (fig. 3 c) is far from a linear partition and presents important fluctuactions in January, February and April.

When tidal amplitude is the highest (April) and when the fresh water inflow is maximum (June), the upper layer follows better the theoretical branching both during ebb and flow. When the tidal amplitudes are the lowest (March and May), the asymmetry seems to be the greatest. Nevertheless, the strong asymmetry of February has no obvious cause. On the contrary, in the mid layer, the asymmetry is great for the highest tides (April) ; however the asymmetries in January and February are obscure.



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salinities, indicates that strong changes of salinity occur with the appearance or disappearance of the jet profile, respectively between the cycles 1-2 and 4-5 ; moreover a landward jet, existing during the seventh cycle, disappears with a decreasing salinity.

The diurnal oscillation is not obvious on the horizontal circulation of fig. 4 ; however it becomes clear on the residual, relative, vertical motion of the maximum vertical salinity gradient. The diurnal oscillation of fig. 6 is still well marked in the maximum velocities during flow.

During October, we observe increasing periods between the appearances of the maximum flow velocities which coincide with the residual seaward anomalies. On the contrary, when the period decreases below that of the semi-diurnal tide, the residual anomaly is landwards in the mid layer. The frequency of maximum ebb velocities remains quasi-constant, with a slight tendency below the semi-diurnal.

In short, those results indicate that oscillations existing in a frequency range lower than the diurnal, may give rise to beats, from which arise the anomalies of circulation. In fact, the diurnal tidal oscillation in August, seems to reduce the lagging and, consequently, the anomalies with respect to the strong anomalies of October.

## ANOMALIES OF VERTICAL MIXING

Is the gradient Richardson number a good indicator of vertical mixing or not ? generally the small tides (March and May) generate values of Ri frequently greater than 2, while high tides give numerous values lower than 2. However, the effects of vertical mixing for the same tides are different in the two branches : a glance at fig. 7 indicates the habitual stronger stratification in the western channel (segments Y are larger than segments B). But, on the whole, the differences between the eastern and western Ri values are not significant.

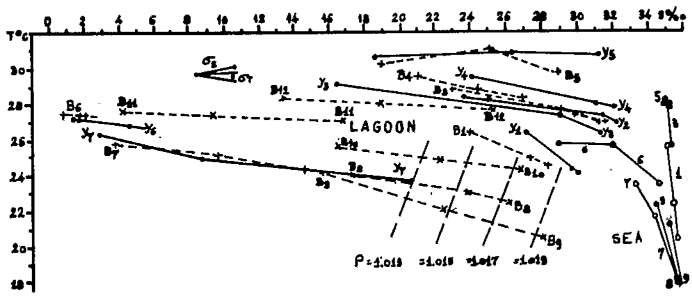


Fig. 7 - Temperature - Salinity - Depth diagrams (numbers indicate months)
+ and . = 2, 4, 6 m depth  o = 0, 10, 20 m sea depth
Lagoon observations averaged on a tidal cycle
$\sigma_S$ and $\sigma_T$ are the standard deviations on 8 successive tidal cycles.

We have pointed out a large time scale variability of velocity profiles in the eastern channel, with the appearance of jet profiles. We observe a decreasing tendency of the Ri values in presence of these anomalous profiles :

TABLE 1

Richardson's numbers at maximum velocities

| | | AUGUST | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|
| tidal cycle n° | | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 |
| FLOOD | upper | 0.5 | 1.3 | 5.4 | 7 | 556 → | 2.5 | 2600 → | 3.7 |
| | lower | 1.8 | 11 | 0.6 | 0.2 | 0.6 | 12 | 0.2 | 4.1 |
| EBB | upper | 59 → | 5.3 | 10 | 1.8 | 0.6 | 58 → | 0.7 | 4.8 |
| | lower | 0.5 | 0.5 | 0.9 | 0.0 | 11 | 1.3 | ∞ | 3.8 |
| averaged Ri | | 15 | 4.5 | 4.2 | 2.2 | 142 | 18 | – | 4.1 |
| | | OCTOBER | | | | | | | |
| tidal cycle n° | | 1 | 2 | 3 | 4 | 5 | 6 | | |
| FLOOD | upper | 49 → | 0.5 | 123 → | 5.3 → | 0.8 | 22 | | |
| | lower | 4.7 | 1.7 | 1.1 | 14 | 4.9 | 0.2 | | |
| EBB | upper | 0.0 | 1.0 | 0.2 | 56 → | 0.0 | 1.5 | | |
| | lower | 0.4 | 0.0 | 0.4 | 7.0 | 6.4 | 1.9 | | |
| averaged Ri | | 13 | 0.8 | 31 | 21 | 3.0 | 6.4 | | |

To the anomalies of cycles 2, 3, 4 in August and cycles 2, 4, 5, 6 in October often correspond averaged Ri significantly lower than in the other events ; more precisely, the process of destabilization between two consecutive cycles occurs in the upper layer, as indicated by the arrows on the table 1.

To our mind, the process of destabilization which appears at frequencies lower than the semi-diurnal tide, contributes moreover to the vertical mixing by changing the conditions of maintenance of turbulence. The ratio Az/Kz may define a critical value of the Richardson number (TAYLOR, 1931, PROUDMAN, p. 101, 1953) below which turbulent energy is supplied from the mean motion. Kz and Az, the coefficients of eddy-diffusion and of eddyviscosityare computed from the residual velocity and salinity fluctuations in August and October ; fluctuations of vertical velocities are estimated from the vertical oscillations of the salinity around its mean value at the depth 3 m. We obtain two different critical values, 2.1 in August and 6.5 in

$\overline{w's'}$ increases tenfold in August with respect to October.
Lastly Kz is found larger in August than in October : their relative magnitude equals 8.

The ratio Az/Kz, estimated at a smaller time-scale, from the temperature and current AANDERAA measurements every ten minutes,gives the following results : during the lowest tidal amplitude (March) its values are 4.6 at location B and 3.0 at location Y ; during the highest tidal amplitude (April) its values are 0.10 and 0.11. For these very different tidal velocities (multiplicative factor of 3) the coefficient of eddy-diffusion Kz is in the range 0.01 - 1 $cm^2\ s^{-1}$ (low amplitude) and 5 - 50 $cm^2\ s^{-1}$ (high amplitude) : the highest value is found at Y for low amplitude, at B for high amplitude.

The principal theoretical and experimental results quoted by WELANDER (1968, p. 22-26) indicate that turbulence can be sustained when the flux RICHARDSON number, defined as $R_f$ = (Kz/Az). Ri, lies generally below the mean value 0.3. From the local Ri observed during March, April, August and October it appears that $R_f$ is about a few unities. Theoretically, and that is observed by comparison of March, August and October, the ratio Az/kz does not depend on the scales. The too large values of the observed $R_f$ arises, to our mind, principally from the vertical scale of the local Ri observed : the vertical gradients should be estimated every 30 cm on the vertical, in order to obtain realistic values of Ri and hence of $R_f$.

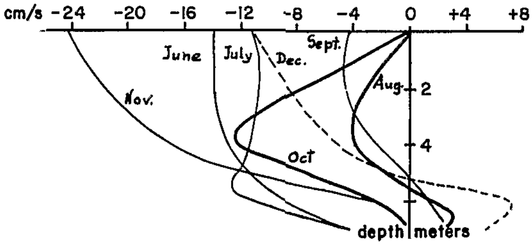


Fig. 8 - Means of the velocity profiles in the eastern channel.

DISCUSSION

The evolution of transports on fig. 8 is coherent with the habitual monthly fresh water inflows (TASTET, 1974) : the transport of about 600 $m^3\ s^{-1}$ in June corresponds well to the strong rainfalls in 1976. The anomalous profiles of August and October may represent transient stages of the river discharges for which the mean wind drift is opposite. The T-S diagrams of January (fig. 7) indicate more mixing than the consecutive months of the dry season : It is well known that atmospheric circulation is particular in January, with a seaward wind which gives rise to a coastal upwelling. COLIN (personal communication, 1977) shows a significant diurnal pike and an important variability around 4-6 days, for the annual wind spectrum in Abidjan. These scales correspond well to the changes of residual salinities and velocities observed in the eastern channel. WEISBERG (1976) demonstrates the effect of the wind variability on the estuarine circulation, and the necessity of measuring numerous tidal cycles, in order to obtain the "mean" circulation. Obviousness, the habitual S W wind has a very different effect on the residual circulation of the eastern and western channels : fig. 9 a indicates that anomalous profiles are often generated in the eastern channel, because the wind drives the circulation landward. On the contrary in the western channel, (fig. 9 b), the seaward circulation is favoured in the upper layer and, consequently, the typical estuarine circulation appears better.

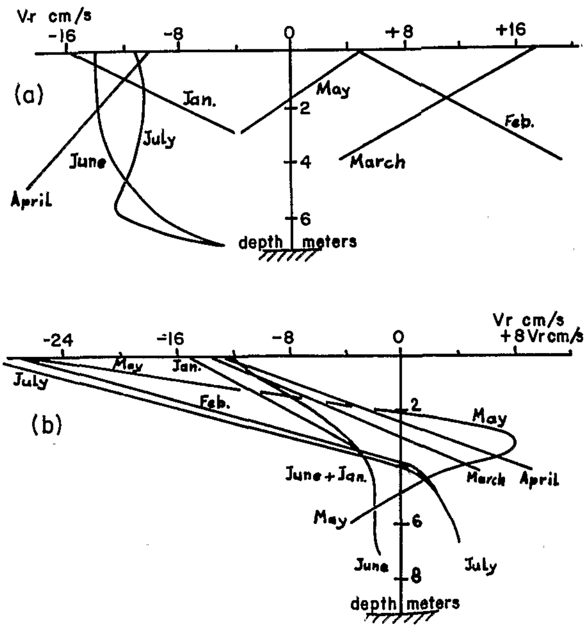


Fig. 9 - a - unsteadiness of the residual velocity profiles at station B.
b - steadiness of the residual velocity profiles at station Y.
(extrapolated from 1 or 2 meters depth to the surface).

CONCLUSION

We have observed a great variability of the residual circulation in a branching lagoon estuary. That variability may give rise to asymmetries between the eastern and western channels. The coefficient of eddy-diffusion Kz presents too a high range of variability which could be estimated from the dimensionless ratio Az/Kz. With respect to the general theoretical and experimental results which give a flux RICHARDSON number in the range 0.1-0.5, it appears that the gradient RICHARDSON numbers should be observed with a vertical distance of about 30 cm. The effect of the wind direction and velocity on the asymmetry is pointed out.

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