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Manche
Pas de Calais
Débit
Modèle
Marée

Channel
Dover Strait
Flow
Model
Tide

Computed residual flow through the Dover Strait

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ABSTRACT

A two-dimensional numerical model was used for the empirical deduction of a mathematical expression of water flow through the Dover Strait, for all conditions of tide, wind and ocean slope.

The theoretical results were correlated with sea measurements of radioactive tracers discharged at La Hague cape.

Nine actual years (1983-1991) were then examined, determining an average annual flow of 114000 m^3s^{-1} with a very high monthly variability, but with little difference from one year to the next.

These residual currents being highly sensitive to weather conditions, the mildness of the past few years has occasioned a decrease in fluxes towards the North Sea.

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RÉSUMÉ

Flux d'eau, calculé, à travers le détroit du Pas-de-Calais

Un modèle numérique bidimensionnel est utilisé pour déduire empiriquement une expression mathématique du flux d'eau dans le Pas de Calais, pour toute condition de marée, de vent et de pente de l'océan.

Ces résultats théoriques sont calés d'après les mesures en mer des traceurs radioactifs rejetés au cap de la Hague.

Neuf années réelles (1983-1991) sont ensuite examinées, dont on déduit un flux moyen annuel de 114000 $m^3 s^{-1}$ avec une très forte variabilité mensuelle, mais peu de différences d'une année à l'autre.

Ces courants résiduels étant très sensibles aux conditions météorologiques, la clémence des quelques années passées, s'est traduite par une décroissance des flux vers la Mer du Nord.

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INTRODUCTION

Modern oceanography attaches particular importance to determining water and matter fluxes, both as a matter of scientific progress and as an element in the management of coastal seas.

From this point of view, straits separating and linking two bodies of water are of particular interest. Currents are more intense there, flow sections are smaller and measurements are taken more easily.

However, the problem remains complex and ocean variability means that depending on the time scale applied, these volumes of water can vary considerably.

For the Dover Strait, the question is not a new one. Since the turn of the century, many attempts have been made using current-meters, salt budgets, chemical tracers, differences of electrical potential between the two shores, or mathematical modelling.

The results have turned out to be rather conflicting, both for the overall assessment evaluated from 55 000 m^3s^{-1} (Van Veen, 1930) to 238 000 m^3s^{-1} (Cartwright, 1961) and for its elementary breakdown due to tides, weather conditions or to the oceanic internal pressure gradient, on a very large scale.

The question is dealt with here through combined use of mathematical models and radioactive tracers.

We attempted to improve the assessment of monthly and annual average flows, while detailing the role played by each of the factors mentioned above. In this way, by improving understanding of the mechanisms, estimations can be made, both in the future and in the past, possibly reaching the required level for management objectives in the Channel and the North Sea.

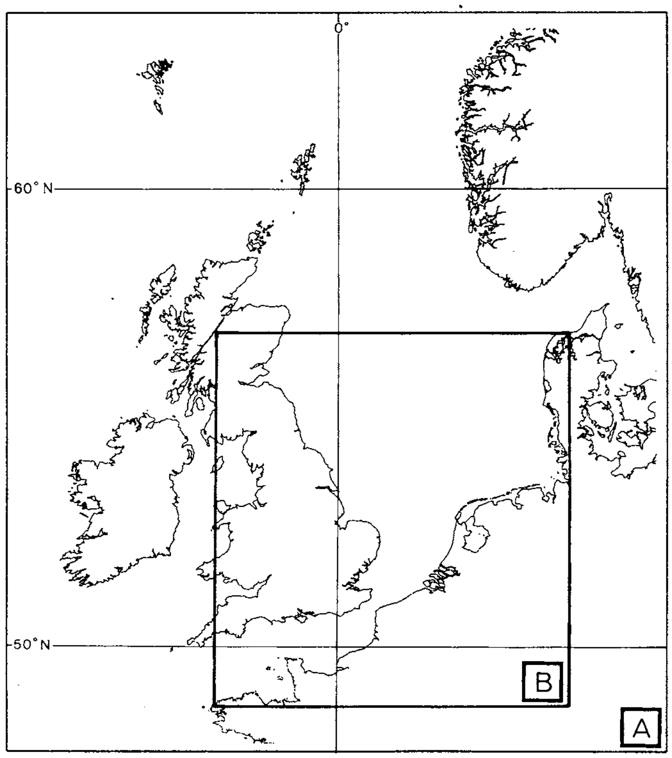

Figure 1

Geographical map showing model boundaries. A: large-scale model; B: submodel used to compute flows through the Dover Strait.

Limites géographiques des modèles. A : modèle général ; B : sous-modèle utilisé pour le calcul des flux dans le Pas-de-Calais.

MATERIALS AND METHODS

The currentology mathematical model and its scope

In order to respect a fundamental modelling principle requiring that a model's borders be set beyond the area of influence of the parameters to be tested, we set the calculation limits outside the continental shelf (Fig. 1).

An initial two-dimensional model was limited to meridians 12°W and 12.5°E, and to parallels 47° and 63°N. The hydrodynamic equations were expressed in spheric coordinates and the resolution was made on a mesh of 10' longitude and 6' latitude (Salomon and Breton, 1990). This model was forced on the edges by tide conditions extracted from Schwiderski's atlas (1983) and over its area by wind stress which we assumed to be spatially uniform and constant.

As this model did not provide a sufficiently discriminating description of the strait region, a submodel was created within it.

In the first phase, a Channel model, its mesh equal to one nautical mile (Salomon and Breton, 1991) was used, but its northern limit, located at 51°20' latitude, proved to be too close to the strait to provide satisfactory results. We therefore introduced a second model of much vaster scope (Fig. 1).

This second model's mesh equals 3 km and stretches between latitudes 48.5° and 57°N and longitudes 4.5°W and 9°E. It solves the Saint-Venant equations in their usual form:

$$\frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + g\frac{\partial \zeta}{\partial x} - fv - e\left(\frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 u}{\partial y^2}\right)$$

$$+ g\frac{u\sqrt{u^2+v^2}}{k^2H^{4/3}} + \frac{\tau_x}{\rho H} = 0$$

$$\frac{\partial v}{\partial t} + u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + g\frac{\partial z}{\partial y} + fu - \varepsilon\left(\frac{\partial^2 v}{\partial x^2} + \frac{\partial^2 v}{\partial y^2}\right)$$

$$+ g\frac{u\sqrt{u^2+v^2}}{k^2H^{4/3}} + \frac{\tau_y}{\rho H} = 0$$

$$\frac{\partial z}{\partial t} + \frac{\partial (Hu)}{\partial x} + \frac{\partial (Hv)}{\partial y} = 0$$

u,v: velocity components
x, y: horizontal coordinates
t: time
H:water depth
ζ: surface level
f: Coriolis factor
k: friction coefficient
ε: horizontal viscosity
ρ: mass per unit volume
τ_x, τ_y: wind stress components.

The numerical method, in finite differences, is of the Alternate Direction Implicit type (A.D.I.). It has been described at length in literature since Peaceman and Rachford (1955) and J.J. Leendertsee (1970).

Main simulation difficulties

An extremely delicate point in our simulations was to ensure almost perfect convergence of model results in a situation which no longer depends at all on the initial calculation conditions.

To assess the error made in case of incomplete stabilization, we can note that a disparity of 1 cm corresponds, over the entire North Sea, to a volume of 4.6 x 10^9 m^3, *i.e.* a flow of 100000 $m^3\ s^{-1}$ during an entire tidal period. Even considering that the compensating flux would not take place in majority through the Dover Strait, but rather *via* open borders of the area, we can consider the required model stability to be about a millimetre.

Faced with the same problem, Pingree and Griffiths (1980) had to stabilize their model during five tidal cycles, as did Prandle (1978 *b*) over four days. Here, we had to extend this phase over twenty tides before fulfilling the criteria mentioned above.

Another difficulty arises from the lack of precision of velocity limit conditions used on the outskirts of the detail model. When a model is steered at its limits by level conditions, currents are assessed using a complementary relation, for instance of the "radiation" type, which does not provide a high degree of precision.

According to the Bernouilli relation:

$$\frac{V_2}{2} + g\zeta = \text{constant}$$

a ΔV error on the level, corresponding to the previous value, multiplied by the g/V factor which is equal to about 10 (M.K.S.). This means that an error of a few centimetres per second in estimating currents along the boundaries would be the equivalent of an error on the level of a few millimetres, with the disturbance effect evaluated above.

This is why we defined a larger submodel than that of the Channel.

External conditions

Both mathematical models described above (general model and submodel) were used for around thirty situations combining the four external parameters: wind strength and direction, tide amplitude and average level of the sea above the continental shelf.

In each case, we obtained a field of residual currents and a flow value in the Dover Strait. We then looked for a simple mathematical formula to express these numerical results as a function of the four external parameters.

Linking to true conditions

The hypotheses of these simulations are particularly oversimplified, especially for weather situations, making them difficult to use with true conditions. Therefore, a verification and calibration phase was desirable.

For this, we proceeded by an indirect method, simulating advection and dilution in the Channel of the radioactive discharges from the nuclear fuel reprocessing plant at La Hague. We used the Lagrange residual current method (Orbi and Salomon, 1988; Salomon *et al.*, 1988; Salomon and Breton, 1991) and applied the simulation to antimony 125, known to be almost perfectly conservative.

By determining issue peaks at their source, and the concentration maxima which correspond to them in the Dover Strait, we were able to verify that transit times, and thus current velocity and water flux were correct.

This method for modelling dispersion in barycentric coordinates has already been described elsewhere (Salomon *et al.*, 1991) and will not be discussed here. With regard to the first study, we additionally took into account a succession of various weather conditions and not merely an average wind situation.

RESULTS

Tide alone

Without any influence of the wind or average ocean slope, the following mathematical expression turned out to be the best:

$$\phi = 37400\ \beta^{2.3} \qquad (1)$$

where ϕ is the volume transport rate of water (in $m^3\ s^{-1}$),

β is the ratio of the amplitude of the tide in question (A) and that of the average tide (A_0),

$\beta = A/A_0$,

β may vary from 0.286 for the weakest neap tides to 1.714 for the strongest spring tides;

This value represents the total flow, summing the three current components which are sometimes calculated separately: Euler's residual velocity, Stokes' drift and Lagrange's drift (Cheng, 1986).

It can be compared to values given in the bibliography for an average tide:

Wyrtki (1952, *in* Postma, 1990): 33 000 $m^3\ s^{-1}$
Cartwright (1961): 50 000 $m^3\ s^{-1}$
Pingree and Maddock (1977): 50 000 $m^3\ s^{-1}$
Prandle (1978 *a* and *b*): 36 000 $m^3\ s^{-1}$
Pingree and Griffiths (1980): 30 000 $m^3\ s^{-1}$
Pingree and Maddock (1985): 40 000 $m^3\ s^{-1}$.

All these positive values are oriented towards the North Sea.

The result obtained here of 37400 $m^3\ s^{-1}$, is higher than our first estimate 27000 $m^3\ s^{-1}$ obtained from the Channel model (Salomon and Breton, 1991) for the already mentioned reason of the northern limit's poor positioning in the previous model.

The dependence of the flow to a power of 2.3 of the tidal amplitude at the model's limits should be correlated to the theoretical expressions of Lagrange's residual velocity (V_{rl}) and the residual transport velocity (V_{rt}) in the simple case of a progressive sine wave.

In that case, a particle tidal trajectory may be approximated, expanding the velocity in Taylor series and integrating it over time. This leads to the following expression :

$$V_{rl} = V_{rE} + \frac{V_0^2}{2C}$$

The transport velocity is obtained by averaging the product of water depth and particle velocity, and dividing it by the average water depth. This gives :

$$V_{rt} = V_{rE} + \frac{\overline{\zeta V}}{H}$$

V_{rE} is Euler's residual velocity, Vo is the tidal current's amplitude. C is the wave's speed, and $\overline{}$ represents the averaging operator on a tidal cycle.

According to these formulae, the additional term to Euler's residual velocity would be quadratic if the tidal currents were proportional to the surface variations, which is not far from our result. However, the simplifying hypotheses which lead to these expressions have not been verified in the Channel and theoretical deductions can only give an indication of scope.

Combination of tide and wind

The friction created by the wind on the surface adds another component to water flow in the strait. But because of the nonlinearity of the equations for movement in shallow waters, these two factors are coupled.

The best simple formula synthesizing our results is as follows:

$$\phi = 37\,400\ \beta^{2.3} + 1.1 \times 10^6\ |\tau|\ \beta^{-0.28} \cos(\alpha - 186) \qquad (2)$$

in which

α is wind direction (in degrees), increasing from the north, clockwise,

τ is wind stress

This additional component for water flux is proportional to the wind stress. The theory explains this by the approximative linearity of the residual movement equation.

The maximum wind efficiency angle (6 and 186°) is almost identical to that determined by Pingree and Griffiths (1980): 7 and 187°, but lower than the value given by Prandle (1978 *b*): around 23 and 203°.

The proportionality coefficient is also very close to that given by Pingree : 1.2×10^6.

The $\beta^{-0.28}$ factor represents coupling with the tide. This corresponds to the fact that the friction coefficient of the residual movement equation is proportional to the tidal current intensity.

In all, it should be noted that the tide has two effects which can be opposed:

- When tide and wind induce a slight water flow (approximately 30000 $m^{-3}\,s^{-1}$) or negative (towards the Channel) the larger the tidal amplitude, the larger the total algebraic flow.

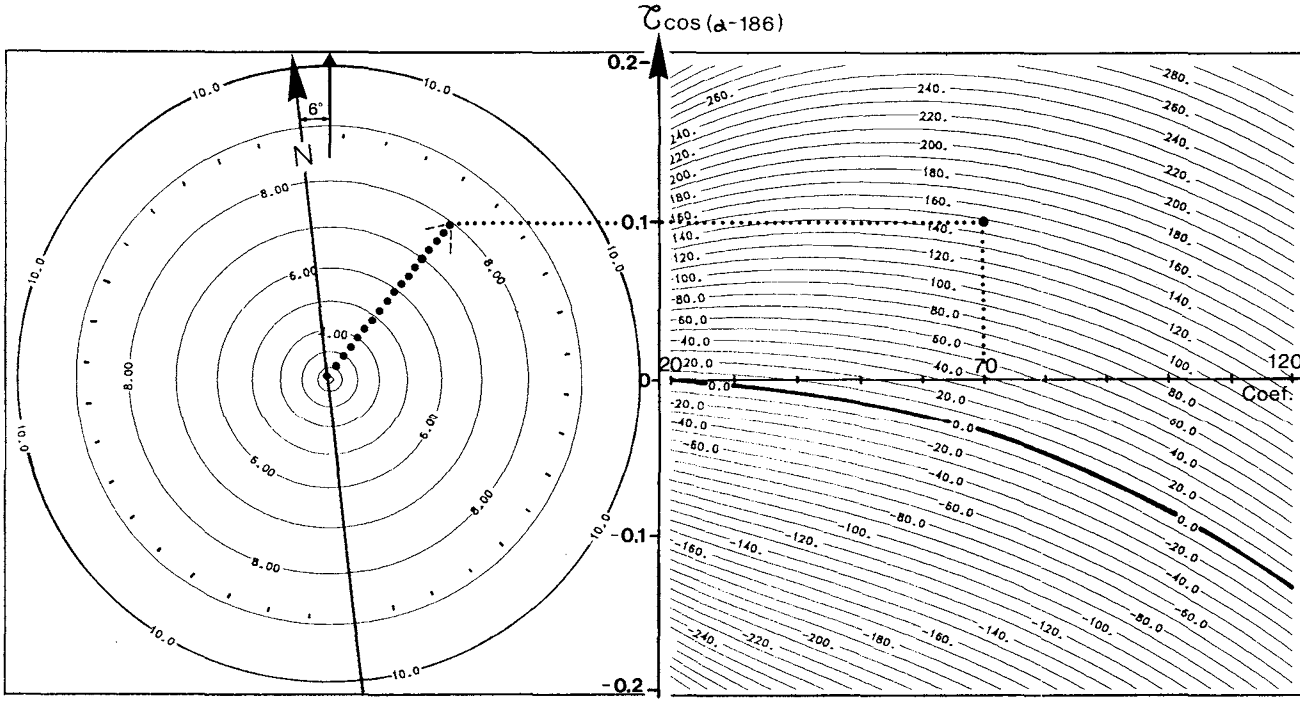

Figure 2

Diagram of equation No. 2 (in units of $10^3\ m^3\ s^{-1}$). Wind stress to be evaluated on the left side, and combined with tidal amplitude on the right. See text for complete explanations.

Abaque correspondant à l'équation n°2 (unité $10^3\ m^3\ s^{-1}$). La tension de vent est déterminée dans la partie gauche du dessin et combinée avec l'amplitude de la marée dans la partie droite (voir texte).

- When tide and wind induce a positive water flow (towards the North Sea) greater than 30 000 $m^3 s^{-1}$, a β_0 value exists for which the total flow is minimal.

$\beta_0 = 1.64\ [\tau \cos(\alpha - 186)]^{0.39}$

If β is less than β_0 and decreases, the wind effect increases as does the total volume transport rate. If β is greater than β_0, and increases, the wind effect decreases, but that of the tide increases as does the total flow (*see* Fig. 2).

Equation (2) is shown as an abacus in Figure 2, expressing the wind stress by the formula:

$\tau = 0.002\ W^2$

In Figure 2, from a given wind intensity and direction, we can construct the "effective" value of the stress (left side of the figure) by projecting the wind vector on the Y-axis. By combining this intermediary result with the β coefficient value (right side of the figure), we directly obtain the flow induced in the Dover Strait by the combined effect of wind and tide. Here, a tidal coefficient equal to β x 70 is used.

As an example, a geometric construction is shown in dotted lines (Fig. 2) for a 8 $m.s^{-1}$ southwesterly wind ($\alpha = 225°$), combined with an average tide. The effective stress obtained on the Y-axis is 0.1 Pa and the resulting flow is about 147 000 $m^3 s^{-1}$.

Average sea level slope

The results of our calculations depend on the possible slope of average sea level on the outskirts of the external model (12°W and 63°N), *i.e.* a difference of levels between the Celtic Sea and the east of the Faroe Islands. We do not have data on the true value of this slope, but the model can determine the relation between a hypothetical value and the water flow component resulting from it.

The model provides a linear response, which was expected, considering the comments made above. The proportionality coefficient equals: 1.4×10^{11}.

$\phi_{\nabla\zeta_0} = 1.4 \times 10^{11} \nabla\zeta_0$

in which $\nabla\zeta_0$ is the average surface slope, in the north-south direction.

As an example, a slope of 5 cm for 1 000 km corresponds to a flow of 7 000 $m^3 s^{-1}$.

For the same slope value, which he considered as reasonable, Prandle (1987) made a similar calculation and obtained a volume transport rate of less than 20 000 $m^3 s^{-1}$. He did not however, provide further details. Therefore, our results are not incompatible, showing that unless we imagine considerable ocean slopes, provoking very violent currents on the shelf break, this term can only be a minor factor in the water volume moving through the Dover Strait.

Lastly, our trials can be summarized by the following formula:

$$\phi = 37400\ \beta^{2.3} + \beta^{-0.28} (1.1 \times 10^6\ |\tau| \cos(\alpha - 186)) + 1.4 \times 10^{11}\ \nabla\zeta_0 \quad (3)$$

Relation to true weather conditions

The above results are relevant for stationary and uniform wind fields, rarely the case in reality. Therefore, we had to examine how far they can be used on a scale of variability in time, and then correlate the true wind and the average wind taken into account in the model.

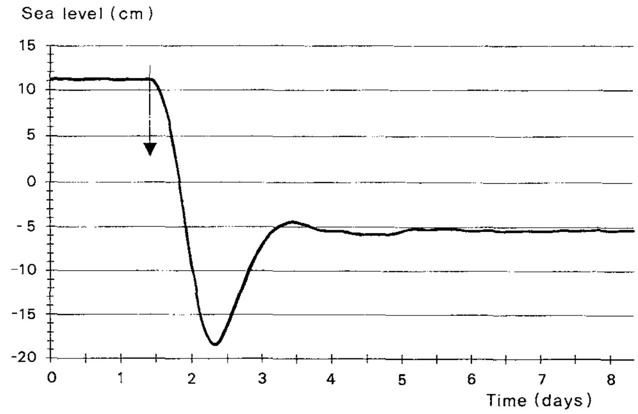

Figure 3

Oscillation of the average sea level following an abrupt wind rotation, as computed by the mathematical model in the southern North Sea.

Oscillations du niveau moyen de la mer après une rotation brutale du vent. Résultat obtenu par modèle mathématique, pour le sud de la mer du Nord.

For the former, we carried out a theoretical simulation considering a wind first orientated north, abruptly shifting south. Figure 3 shows the evolution of the average sea level at a given point, south of the North Sea.

We detected an associated water level oscillation period of two days, and a decay period of four days. This simulation showed that the system memory with regard to weather forcing is limited to four days, and that real situations should be described using averages over this period.

On this scale, the spatial variability of wind fields is much less than for instant fields (*see* Hohn, 1973, for example) but they remain geographically variable, thus the introduction of a correction term for average wind fields will probably be required.

On the second point, as mentioned above, we calculated dispersion of antimony 125 discharged from La Hague cape in the Channel and the North Sea. Some peaks can be identified over a period of approximately sixteen months, until they reach the Baltic Sea entrance (Salomon *et al.*, 1991; Guéguéniat *et al.*, communication to the Channel Symposium). This exercise was repeated over nine years, from 1983 to 1991.

The best correlation between measurements and models, without taking the average oceanic slope into account, was obtained by adding a corrective component of 1.8 $m.s^{-1}$ from south to the average quadratic winds measured at the La Hague semaphore.

Figure 4 provides an example of results obtained at the centre of the Dover Strait, where both good correlation of activity levels and good synchronization of the two signals were found.

This correction does not seem unrealistic. It can be attributed both to loss of variability when averaging winds, to the difference between dominant winds on land and dominant winds at sea, to the difference between average winds in the Channel and the North Sea, or to the average curve of isobars in the region (Hohn, 1973).

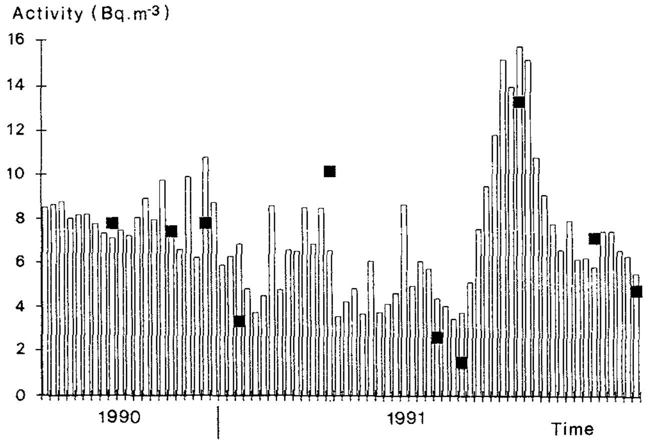

Figure 4

Compared results of calculation from a dispersion model (histogram) and measurement (dots) of ^{125}Sb at strait's centre.

Comparaison des résultats du modèle mathématique (histogramme) et des mesures (points) relatifs à ^{125}Sb au centre du détroit.

It is also possible that the sea friction coefficient was underestimated in the region, due to the tidal current's effect on surface roughness, for example.

Part of this correction should still be attributed to the average surface slope, which we considered as nil, or to a general water density gradient through the North Sea and the Channel.

Application from 1983 to 1991

Monthly fluxes were calculated using equation (2), applied to quadratic winds averaged twice a week, at La Hague, corrected as described above.

The Table presents these results, along with the average monthly values, their standard deviation and average yearly values.

Large differences are observed, with results varying from - 8 000 $m^3 s^{-1}$ (flow towards Channel) in December 1988 to + 288 000 $m^3 s^{-1}$ in December 1989. The fact that these extreme values both occur at the same time of year is a good illustration of flux variability, especially in winter.

The average flow during this period was 114 000 $m^3 s^{-1}$. The highest values occur from October to January and the lowest from May to August.

Prandle (1978 *a*) made the same calculation for the period 1949-1972 and obtained quite similar results in terms of variability and annual evolution, but his results were approximately 30 % higher than ours overall. His extreme values were respectively - 15 000 $m^3 s^{-1}$ and + 364 000 $m^3 s^{-1}$ for an average value of 155 000 $m^3 s^{-1}$. He also observed fairly good regularity in the yearly averages, varying 17 %. In our study, these vary only 13 %.

Figure 5 presents the yearly averages over the past nine years, and seems to reveal a decrease in these flows. The period analysed is too short to make a general observation valid, but the higher values given by Prandle for an earlier period could also correspond to the same phenomenon. It would be particularly interesting to verify this observation over a longer period, since the hydrodynamic functioning of the Dover Strait is closely linked to weather conditions and this could be a sign of global climatic changes.

The calculated fluxes can also be given in terms of transit time, *i. e.* in water mass renewal time: the water volume of the Channel from the Roscoff meridian to the strait is approximately 4 x 10^{12} m^3, thus representing about 15.4 months of flow in the Dover Strait. The same calculation limited to the meridian of Cherbourg gives a water volume of 1.2 x 10^{12} m^3, *i.e.* an average transit time of four months which has most likely evolved over the past few years from two months (December 1987-January 1988) to more than six months (spring-summer 1984 or 1989).

CONCLUSION

Using a numerical hydrodynamic model enables a simple mathematical formula to be empirically drawn up, expressing water flux in the Dover Strait on time scales greater than several days.

These results are generally within the ranges of previous assessments. They refine and complete them by detailing the role played by tidal amplitude and its coupling with wind or average ocean slope.

However, these simplifying hypotheses require that measurement be taken to link the calculation to true conditions. For this, radiotracers have proved to be a powerful tool, enabling the model to be precise to about 10000 $m^3 s^{-1}$, *i.e.* about 1 % of the tidal flow.

The mathematical expression proposed could be used hereafter for any period of interest.

Analysis over the past nine years has revealed an average value of 114 000 $m^3 s^{-1}$, only one third of which was due to tide. On a monthly scale this value varies greatly, and can even be inverted, yet on a yearly scale it is relatively constant.

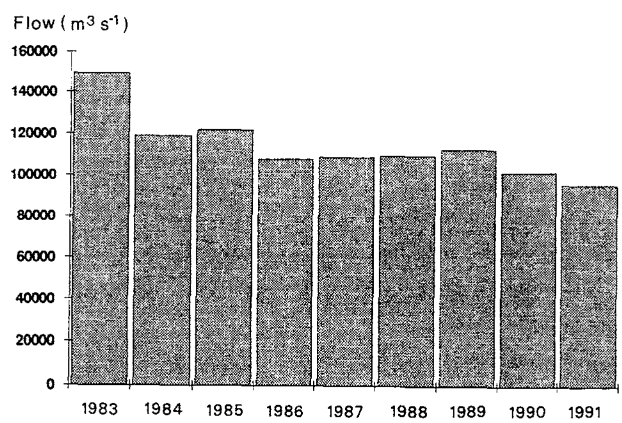

Figure 5

Average annual volume transport rates from 1983 to 1991.

Débits moyens annuels de 1983 à 1991.

Table

Residual flow through the Strait from 1983 to 1991, in units of 10^3 m^3s^{-1}.

Débits résiduels dans le détroit de 1983 à 1991 (moyennes mensuelles - unités 10^3 m^3s^{-1}).

YEAR	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	MEAN
1983	206	58	129	204	220	93	82	65	217	230	87	206	150
1984	232	83	75	63	37	93	87	89	121	168	184	194	119
1985	111	105	99	100	74	125	130	178	127	184	80	150	122
1986	141	8	134	87	155	91	102	134	66	97	157	120	108
1987	58	135	132	124	35	132	52	67	156	142	53	228	109
1988	248	134	55	145	115	90	129	63	77	182	94	- 8	110
1989	150	140	106	61	85	59	57	77	67	138	127	289	113
1990	165	178	111	53	74	97	93	84	79	109	98	80	102
1991	139	75	100	74	35	124	116	88	81	83	134	102	96
MEAN	161	102	105	101	92	100	94	94	110	148	113	151	114
S.D.	60	52	27	49	62	23	28	38	51	48	41	89	15.5

During this period, we believe that a decreasing trend in the flow was indicated. This phenomenon should be studied more attentively, since it could have major consequences for the Channel and North Sea ecosystems.

Acknowledgements

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