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Mechanisms of water exchange at the platform edge of oceanic islands

J.L. Pelegrí, A. Antoranz, C. Gordo, A. Hernández-Guerra, A. Martínez & P. Sangrà
Departamento de Física, Facultad de Ciencias del Mar, Universidad de Las Palmas de Gran Canaria, Spain

Abstract

Islands in deep oceanic waters have rather vulnerable ecosystems due both to the narrowness of their coastal platforms and the presence of very steep slopes. We examine the different mechanisms capable of exchanging water at the platform edge of oceanic islands, and conclude that the principal ones are turbulent transport in the boundary layers generated by intense deep ocean currents and wind-induced cross-slope transport at selected coastal sites.

1 Introduction

A common acquiescence is that coastal areas in oceanic islands are in good health because whatever is thrown to the water is carried to the deep ocean by the currents. The only "evidence" behind this puerile idea, however, is that xenobiotic compounds are not visible along other coastlines or under the sea surface, but this is simply because there are not such coastlines and the water bottom is not discernible. The truth is that coastal waters of oceanic islands are rather isolated and this makes their coastal ecosystems very vulnerable.

Islands in oceanic deep waters are, up to a high degree, separated from the deep ocean waters because of the narrowness of their coastal platforms and the presence of very steep slopes. On one hand the narrowness of oceanic island platforms causes that the condition of non-normal flux at the coast is still rather valid at the shelf edge. On the other hand, the very steep slope of these islands prevents permanent currents to cross the isobaths at the shelf edge. Both factors inhibit water exchange between the coastal and the deep

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oceans, resulting in slow water renewal at the platform edge and undesirably high residence time of these waters.

The main physical processes responsible for the circulation in the coastal ocean and, in particular, for water exchange at the platform edge of oceanic islands, are tides, river discharge, deep-water forcing, and winds. Tidal currents, however, because of their periodic character and orientation almost parallel to the coast, usually cannot provide such exchange. Only internal tidal waves, generated at the slope, may propagate into the platform. River discharge creates a water plume that moves along the coast as a Kelvin wave but, unless the discharge is very large, will hardly penetrate into deep waters.

Turbulence generated near the islands is a very important mechanism for water exchange. This occurs in islands found on the path of relatively strong oceanic currents, characterized by large Reynolds numbers: water import into the platform is then associated to mesoscalar structures generated at the lee of the island, and water export usually takes place further upstream. The residence time of platform waters is linked to the period of generation of the mesoscalar structures, typically several days. In this work we will illustrate this type of situations with data from the island of Gran Canaria.

For islands in the verge of rather weak permanent currents, exchange is essentially controlled by wind. If the platform is very narrow the condition of non-normal flux at the coast still holds at the platform edge and exchange is highly restricted. If the water column is stratified an upwelling-type vertical circulation cell could develop, which needs not to verify an integrated non-normal flux condition. For many small islands, however, upwelling is not significant. Even more, the water column in many instances is well mixed and the velocities show little depth dependence. In this case the pattern of water exchange, as well as the width of the boundary layer between the coastal and deep oceans, may be investigated using Csanady's [1] model of the arrested topographic wave. This model suggests the existence of horizontal cells of recirculation and determines the natural lengths of such cells. In practice, however, the coastal morphology will largely control the generation of coastal filaments that stretch out into the deep ocean.

Our aim here is to examine the limited pathways for water, and other substances, to be exchanged at the platform edge of oceanic islands. The survival of coastal ecosystems in these islands depends, first of all, on our capacity to understand and quantify how this exchange takes place, and second, but not less important, on our skill to convince the local administration that the island platforms are not an unlimited dumping site.

2 Tidal currents on the island platform

Tidal currents are very susceptible to the presence of continental masses, with the tidal ellipses orientating grossly parallel to the predominant orientation of the coastline up to several hundred kilometers away [2]. The oceanic islands, because of their relatively small size as compared with the

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wavelength of the tide, have little effect on the surrounding deep waters but do control the characteristics of the tidal currents on the platform. Figure 1 illustrates the bathymetry around Gran Canaria and indicates the locations of stations where currents were simultaneously measured, and Table 1 presents the tidal ellipse information for the M_2 component obtained from these current-meters [3]. The results clearly illustrate how the current orientates parallel to the coast over the island platform.

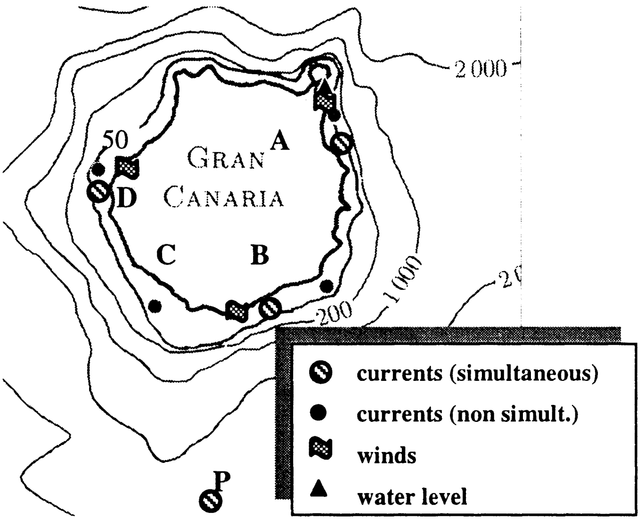

Figure 1. Stations and bathymetry.

Table 1. Semi-major axis, *M*, semi-minor axis, *m*, orientation relative to the east, *O*, and Greenwich phase lag, *G*, for the M_2 ellipses.

Current-meter	*M* (cm/s)	*m* (cm/s)	*O* (°)	*G* (°)
A	9.6	-0.2	143.9	265.7
B	3.78	-0.4	178.5	313.8
C	2.9	0.1	140.3	127.6
D	26.91	1.69	7.42	314.9

The modelization of tidal currents in the presence of topography verifies the importance of the coastal morphology. Figure 2 shows the axes of the M_2 ellipses around Gran Canaria, as obtained from a numerical model using real bathymetry [4]. The currents are always approximately parallel to the coastline over the platform and flow along the isobaths over the slope.

So far we have considered the barotropic tidal wave. A different situation arises with the internal tidal wave, which has a wavelength of only a few

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kilometers, depending on the vertical stratification. The barotropic tidal wave, when running into the coastal slope, generates an internal tide that is either reflected back into the deep ocean or forth into the platform [5]. In this last case there will be cross-slope water movements which, when associated to the different along-shore regimes in the coastal and deep-ocean, will cause permanent exchange. This may be important in those regions exposed to the incoming tidal wave, e.g., along the southern portion of Gran Canaria. This is the situation illustrated in Figure 3, with a rather strong cross-slope tidal current south of Gran Canaria at the 100 m isobath [6].

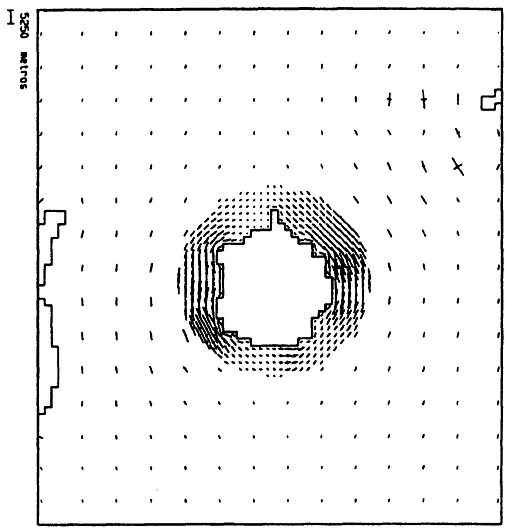

Figure 2. Numerical simulation, using real bathymetry, of the M_2 ellipses around Gran Canaria.

3 The insulating effect of the island slope

Let us consider the simplest possible model, a homogeneous ocean in approximate geostrophic balance, and consider the current flowing into a region where the water depth shows substantial horizontal changes (as compared with changes in the latitude):

$$fv = g\frac{\partial \eta}{\partial x}, \tag{1}$$

$$fu = -g\frac{\partial \eta}{\partial y}, \tag{2}$$

$$\frac{\partial (uh)}{\partial x} + \frac{\partial (vh)}{\partial y} = 0, \tag{3}$$

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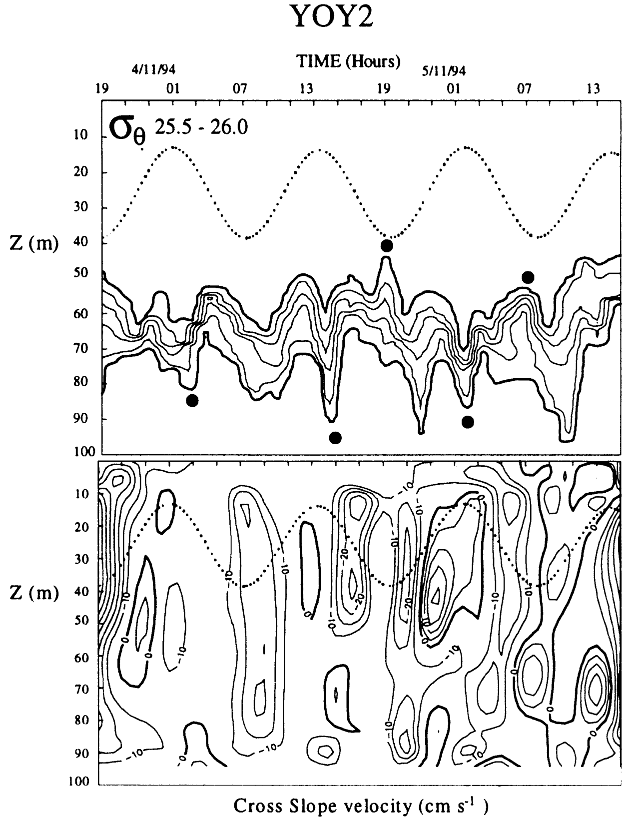

Figure 3. Density stratification and cross-slope velocities near the shelf edge south of Gran Canaria, in the top figure the dotted line illustrates the barotropic tide.

with the usual notation, η being the surface elevation, h the water depth, and x, y the across-shore and along-shore coordinates. It is easy to show that the current will flow along isolines of constant surface elevation, which coincide with the isobaths. Cross diferentiation of eqns (1) and (2) and substitution into (3) leads to the desired condition $\vec{u} \cdot \nabla h = 0$. The deep currents will never flow over a sea mountain or penetrate into the slope of an oceanic island, what we may call the fundamental insulating effect of a steep slope. The presence of horizontal stratification can break this constraint, i.e., if the ocean is stratified along-shore then there may be a geostrophic cross-slope flux. However, stratification is usually normal to the shelf edge [7] and the dominant geostrophic flow is along-slope.

Notice that it is not really necessary to assume geostrophy to derive the condition of along-isobath flow. Consider the case of a homogeneous coastal

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ocean, with the isobaths parallel to a long straight coast. Then the coastal constraint $U \equiv uh = 0$ at the coast, together with eqn (3), implies $V \equiv vh = V(x)$, so that between upstream and downstream transects V remains constant along isobaths, the isobaths behaving as streamlines.

For real applications the model should include other effects such as horizontal stratification, wind stress, and bottom friction acting on the along-slope direction. Csanady and Shaw [8] considered this problem and found that even in this case there is no significant bottom flow escaping from the platform. We will see below, however, that the presence of wind may indeed produce depth-integrated cross-slope currents different from zero, providing a mechanism of water transfer out of the shelf at localized sites.

4 Islands in the path of strong currents

The transition to turbulence in homogeneous flows is assessed by means of the Reynolds number, which compares the stabilizing effect of viscosity against the destabilizing role of inertia. The Reynolds number, $Re = UL/\nu$, is expressed in terms of the horizontal and velocity scales of the flow, L and U, and the horizontal viscosity ν. A classical problem in fluid mechanics is the development of turbulence in a flow past a circular cylinder. For $Re<4$ the flow is laminar and perfectly conforms to the cylinder. For $4<Re<40$ there is a wake behind the cylinder, characterized by two rotating vortices, over which the main stream flows. For $40<Re<200$ the wake becomes unstable and detaches from the cylinder as an oscillation in space and time. As Re increases the oscillations turn into two row of vortices, periodically shed from the two sides of the cylinder and rotating with opposite senses, which has been called a Karman vortex street. When $Re>200$ the vortex street becomes unstable and irregular, and the flow within the vortices chaotic. For Re beyond 5000 the periodicity is lost and the flow is fully turbulent.

The geophysical problem of an oceanic island immersed in the path of a strong current resembles the Karman street vortex problem, aside of the important role played by the earth's rotation [9]. In this case L is given by the size of the island, U by the velocity of the incoming current, and ν by the characteristic oceanic value of horizontal viscosity. As in the Karman street vortex problem, the characteristics of the leeward flow depend on Re. Hence, given L and ν, the leeward flow depends on the impinging velocity U.

Table 2. Relation between U, Re, and the vortex shedding period.

U (m s^{-1})	Re	T (days)
0.04	20	no shedding
0.07	40	no shedding
0.1	60	36
0.19	100	16.2
0.47	250	5.4

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Table 2 summarizes the numerical experiments carried out by Sangrà [9] for a range of realistic flows impinging onto Gran Canaria. The Reynolds number is calculated using ν =100 m^2 s^{-1}, L=54 km, and the value of the impinging velocity for each numerical run. The vortex shedding period T is obtained from the fluctuations of the drag coefficient on the island.

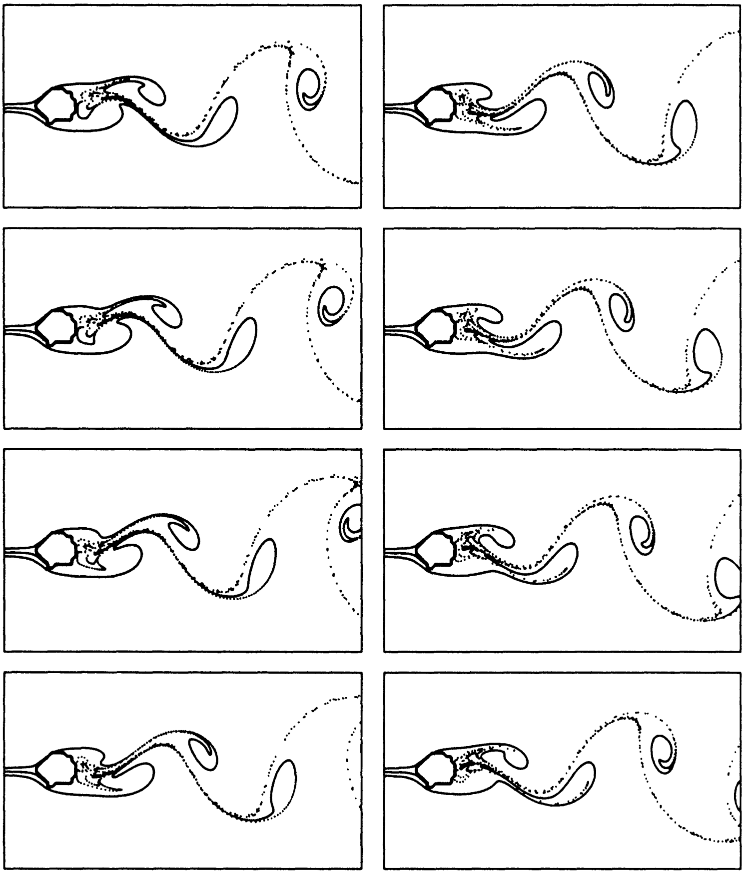

Figure 4. Sequence of tracer lines over one full period for a flow with Re=100 impinging onto a near-circular island which resembles Gran Canaria.

Figure 4 illustrates a full cycle of streak lines using Sangrà's [9] model for Re=100. The sequence goes from top to bottom, starting with the left column, each box separated by time intervals of 1/8 of a period. The flow enters the domain from the left (northeast of Gran Canaria) and leaves it at the right (southwest of Gran Canaria), the shedding of successive cyclonic and anticyclonic vortices is very clear. In this simulation we have chosen

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three streak lines, originated at the upstream end of the domain, which pass very close to the island. The dot points illustrate the position of different particles that have passed through the origin of the streak line at successive time intervals. Where the dot points become closer it means that the current slows down, where they separate apart it means that the current speeds up.

One very important feature is the separation of the flow associated to small capes at both sides of the island, particularly by the one in the lower part (northwest of the island). Another remarkable aspect is the reincorporation of flow, from vortices sometimes quite away from the island, at the lee side of the island. These results are confirmed with satellite images from the region, such as the one in Figure 5 that shows the surface temperature distribution around the island on August 31st, 1999.

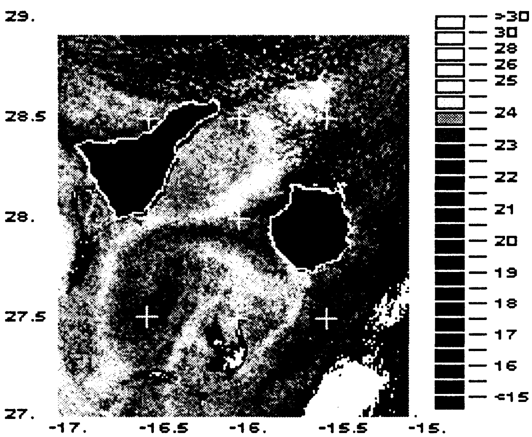

Figure 5. Surface temperature distribution around Gran Canaria on August 31st, 1999.

5 Wind induced exchange

We will briefly examine now whether wind may provide a mechanism for water exchange across the shelf edge of oceanic islands. The coastal constraint, together with the vertical homogeneity of the water over the shelf, inhibits the existence of significant cross-shelf flow near the coastline. In the far deep ocean, however, the surface Ekman transport is perpendicular to the wind. Hence, if we have a situation where the wind blows parallel to the coast we need to understand how, and over which distance, the transition

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between the coastal constraint and the Ekman flux takes place. This problem was examined by Csanady [1]. Depth-integration of the eqns (1) and (2), including the surface and bottom stresses, gives:

$$fV = gh\frac{\partial\eta}{\partial x} + F_x, \tag{4}$$

$$fU = -gh\frac{\partial\eta}{\partial y} + F_y - B_y. \tag{5}$$

F_x and F_y are the wind-stress components, $B_y \equiv -rv$ is the along-slope bottom stress, inversely proportional to the velocity with a friction coefficient r typically of about 10^{-4} m s^{-1} [10], and the across-slope bottom stress is neglected because of the low cross-shelf velocity.

By taking the curl of eqns (4) and (5), with $h=h(x)$, the following depth-integrated vorticity tendency balance is obtained:

$$\frac{\partial^2\eta}{\partial x^2} + \frac{f}{r}\frac{dh}{dx}\frac{\partial\eta}{\partial y} = \frac{f}{rg}\left(\frac{\partial F_y}{\partial x} - \frac{\partial F_x}{\partial y}\right). \tag{6}$$

The solution of eqn (6), subject to appropriate boundary conditions, determines the transition between the coastal constraint and the Ekman offshore transport. Consider the case of constant wind, parallel to the coast, which makes the right hand side of eqn (6) zero. The coastal constraint and eqn (5) imply $F_y = rv$ at x=0, which using (4) becomes the first boundary condition for (6), $\partial\eta/\partial x=(fF)/(rg)$at x=0. A second boundary condition corresponds to zero surface elevation far away from the coast, η=0 as $x\to\infty$.

Under these conditions eqn (6) has the form of the heat conduction equation in one dimension, with surface elevation playing the role of temperature and negative y the role of time. The longshore velocity, proportional to the elevation gradient, is the analog of heat flux, so the first boundary condition by the along-shore wind is like prescribing a heat flux at the coast. The boundary condition at large x is like requiring a constant temperature far from the end of the semi-infinite slab. Csanady [1,11] solved this equation for different wind forcing or initial surface distributions. He noted that surface perturbations, such as those produced by river discharge, propagate along the shelf in the same direction as Kelvin waves. He also found that wind forcing is responsible of exchange at rather localized positions of the shelf-edge. In real cases, where the coastal morphology is quite different from an idealized straight coast, it seem reasonable to expect that the exchange at the shelf edge will be related to features such as capes.

We may estimate the distance over which the coastal constraint vanishes by simply asking the two terms on the left-hand side of eqn (6) to be of the same order, as $l_x \approx (rl_y / f\alpha)^{1/2}$. This means that the size of the island, characterized by l_y , controls the width l_x over which the flow adapts to the deep-ocean Ekman transport. For small islands, say about 10 km long, this width is going to be approximately 1km. This suggests that small islands with narrow shelves will have very limited exchange at the shelf-edge.

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