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ISLAND WAKES

E. D. Barton, University of Wales, Bangor, Menai Bridge, Anglesey, UK

doi:10.1006/rwos.2001.0140

Introduction

The ‘island mass effect’ has been documented for about half a century. This refers to a biological enrichment around oceanic islands in comparison to surrounding waters. Despite a relatively large body of evidence to support the existence of such an effect in the vicinity of islands, there have been few studies to investigate the underlying physical causes of this phenomenon. From a physical point of view the presence of an island in the background flow will disturb the flow regime to produce perturbations that ultimately must have biological consequences.

Two types of island disturbance have been investigated. The first takes place in shallow, stratified shelf seas with significant tidal regimes but no appreciable mean flow, where as the tide moves water back and forth the island acts as a stirring rod to enhance vertical mixing locally and break down the pycnocline. The second occurs in both shallow and deep water, where flow past an island generates eddies downstream and a wake of disturbed flow extends several island diameters away. This arises in the case of larger islands when a clear ambient flow dominates over tidal variability and also around islands small enough that the tidal stream itself can generate a similar effect. The scale of the eddies is typically close to the island diameter and their time scale will be several days for larger islands but only hours in the case of tidal flows. The nature of the wake and eddies may differ between the oceanic case where conditions can be considered quasi-geostrophic and the shallow case where they are frictionally dominated by bottom stress.

Theory

Nonrotating Case

The simplest case is where flow past isolated oceanic islands is considered to be analogous to that of channel flow past a circular cylinder. The work of Batchelor presented the case of nonrotating flow in a homogeneous fluid. The form of the downstream disturbance or wake is related to the value of the Reynolds number

$$R_e = \frac{Ud}{\nu} \qquad [1]$$

where U is the free velocity upstream, d is the diameter of the cylinder and ν is the molecular viscosity. Although this formula appears simple, there are certain practical difficulties in applying it to even a nonrotating ocean of homogeneous character. Few islands are isolated or cylindrical, the upstream velocity is generally not well known and finally the molecular viscosity must be replaced by the horizontal eddy viscosity in the ocean, which is in general poorly known. Studies of Aldabra, an Indian Ocean atoll, indicate that the same current speed impinging on the island from different directions produces a different wake because of the asymmetrical form of the island. It is frequently the case that the upstream current in the ocean varies on a range of time scales, or may be subject to horizontal shear, both of which complicate the choice of a suitable value for the free stream velocity. Values of horizontal eddy viscosity coefficients in the ocean based on many experimental determinations increase with the length scale of interest, l, in a nonlinear fashion $K_h\,(\mathrm{m}^2\,\mathrm{s}^{-1}) = 2.2 \times 10^{-4}\, l^{1.13}$. Typical values appropriate to oceanic islands vary between 10^2 and $10^5\,\mathrm{m}^2\,\mathrm{s}^{-1}$.

The nature of the wake downstream of the obstacle changes as the Reynolds number increases (**Figure 1**). For low Reynolds numbers ($R_e < 1$) the flow pattern downstream is the same as that upstream and there is no perceptible wake. The flow remains attached to the sides of the cylinder and is laminar throughout the flow field. At Reynolds numbers between 1 and 40, the wake remains basically laminar away from the cylinder and two eddies are formed immediately behind the obstacle where they remain attached. At higher Reynolds number the wake becomes increasingly unstable and counter-rotating eddies form a vortex street. The eddies expand as they move away from the obstacle and gradually decay. For Reynolds numbers $R_e > 80$ eddies formed behind the island no longer remain trapped but are shed alternately into the vortex street. The frequency of eddy shedding n is related to another nondimensional number, the Strouhal number

$$S_t = \frac{nd}{U} \qquad [2]$$

This number approaches an asymptotic value of 0.21 at higher Reynolds numbers.

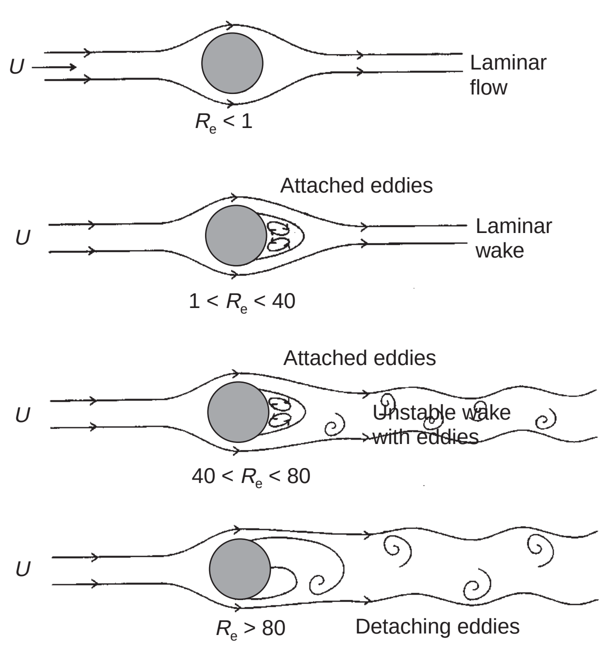

Figure 1 Reynold number regimes for flow past a cylinder.

Effect of Rotation

The ocean of course is on the rotating Earth and so laboratory experiments have been carried out in rotating tanks to investigate the effect of this rotation. Two dimensionless quantities of importance here are the Rossby number

$$R_0 = \frac{U}{\Omega d} \qquad [3]$$

which represents the importance of rotation at rate Ω and the Ekman number

$$E_k = \frac{2\nu}{\Omega d^2} \qquad [4]$$

which determines the width of the wake. The ratio of Rossby number to Ekman number is proportional to the Reynolds number, generalizing this concept to the case of rotating flow. The Earth's rotation enhances the shedding of eddies in the same sense (cyclonic) so that in the Northern Hemisphere predominantly anticlockwise eddies are to be expected whereas in the Southern Hemisphere clockwise should be more common.

Because of the spherical shape of the Earth's surface the rate of rotation about the local vertical varies with latitude $f = 2\omega \sin\phi$, where ω is the rotation rate of the earth and ϕ the latitude. This variation of the Coriolis parameter f can be represented in terms of the 'beta plane' where $\beta = \mathrm{d}f/\mathrm{d}y$ and y is distance north of the equator. If the dimensionless parameter

$$\beta' = \frac{\beta d^2}{4U} \qquad [5]$$

is small, the beta effect may be ignored and the rotation rate taken to be constant on the scale of the island in question. If on the other hand it is of order unity, differences occur from the case of uniform rotation. In particular flow separation is enhanced for flow towards the west and inhibited for flow towards the east.

Effects of Bottom Friction

Studies of flow patterns around small islands only a few kilometers in diameter in shallow shelf seas indicate that the Reynolds number as defined in eqn [1] overestimates the value at transition between the different cases of wake formation. It has been found that the island wake parameter

$$P = \frac{Uh^2}{K_z d} \qquad [6]$$

where U is the stream velocity, h is the water depth, K_z is the vertical eddy diffusivity and d is the dimension of the island, provided better agreement than the Reynolds number between observed and predicted wake parameters. However, P is actually a correct formulation for the Reynolds number when the effect of lateral and bottom frictional boundary layers is taken into account.

Observations

Until recently observations of island flow effects had been limited mainly to remote sensing reports of eddy production. Attempts to observe vortex production and development *in situ* behind oceanic islands have been largely unsuccessful because the background flow has been too weak or variable or the methods of observation insufficient to determine the flow regime adequately. However, a classic early report in 1972, based upon sparse observations of the drift of fishing gear and rudimentary surface current measurement, showed drift patterns downstream of Johnston Atoll in the Pacific Ocean in good agreement with a vortex street situation.

In the case of Aldabra Atoll, situated in the South Equatorial Current of the Indian Ocean, surveys made with acoustic Doppler current profiler indicated a single cyclonic (anticlockwise) eddy trapped behind the island in a low Reynolds number regime on two occasions. There was no evidence of continuous eddy or wake production downstream. The flow was variable during the experiment. The first trapped eddy was observed during westward background flow around $20\,\mathrm{cm\,s^{-1}}$. Two subsequent rapid surveys, about ten days apart, of the current field showed predominantly northward and westward flows, respectively. In the first case the free stream flow was about $20\,\mathrm{cm\,s^{-1}}$ and impinged on the island's widest cross-section. No eddy was observed, only an asymmetrical deviation of currents behind the island. The later case found slightly stronger ($30\,\mathrm{cm\,s^{-1}}$) free stream velocity from the east impinging the narrow aspect of the island. This time a second weak eddy of diameter similar to the island width was indicated.

Another island showing highly variably flow regimes is Barbados. In Spring 1991, the flow seemed topographically steered around the Barbados ridge and there was no clear evidence of eddy production. The following spring, anticyclonic and cyclonic eddies of similar size to the island were found on either flank (**Figure 2**). Computer simulations of the

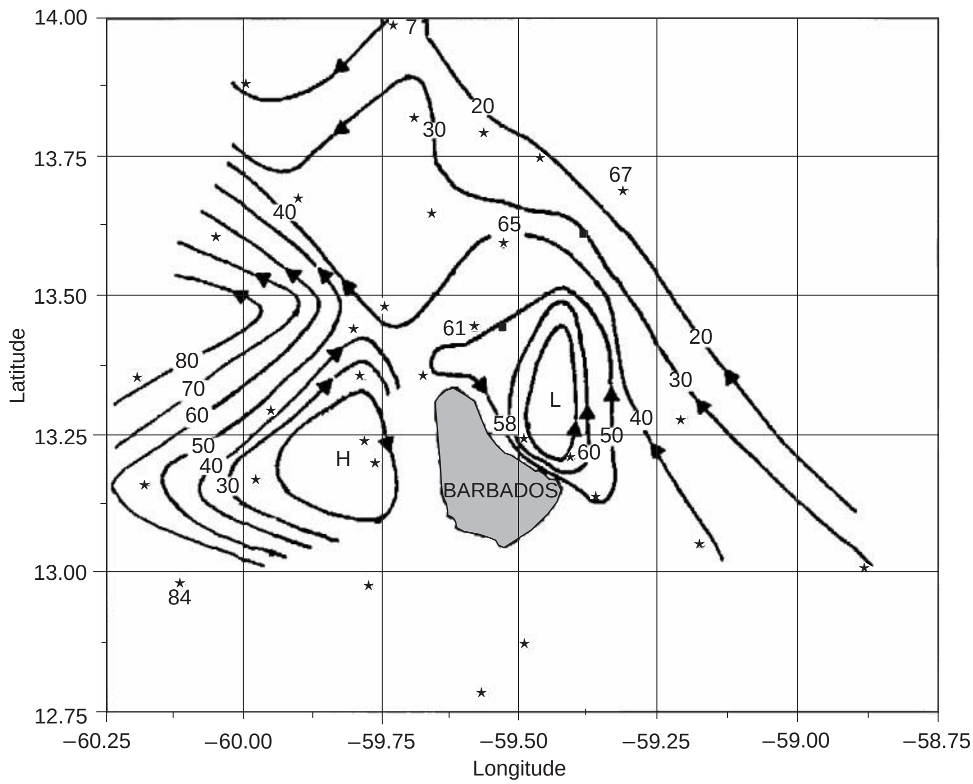

Figure 2 Sea surface topography in centimeters relative to 250 dbar 25 April–2 May 1991. Arrows denote the direction of surface flow. Note the overall northwestward flow and anticyclonic (H) and cyclonic (L) eddies either side of the island. (Adapted from Bowman *et al.*, 1996.)

flow regime indicated that typical conditions were conducive to shedding of cyclones and anticyclones alternately with a period of around 10 days. It was not possible to demonstrate the degree to which the observations were attributable to this type of vortex generation, however, and it was concluded that much more extensive observations would be needed to do so. One interesting indication of the simulation was that there was a continuous region of downstream reverse flow towards the island that could provide a return path for fish larvae swept away from the island.

Considerable evidence of recurrent eddy shedding has been reported recently in the Canary Island archipelago. There, the Canary Current flows southwestward at an average speed of $5\,\mathrm{cm\,s^{-1}}$. Eddies of both signs have been reported (**Figure 3**) as being frequently spun off from the island of Gran Canaria. Cyclonic eddies of the same diameter as the island (50 km) and rotation period around 3 days have been observed to develop on the southwestern flank of the island and to move southwest at speeds of $5–15\,\mathrm{cm\,s^{-1}}$. Almost as frequently, anticyclonic eddies have been observed to develop on the south east of the island. They have diameters up to twice the size of the cyclonic eddies, similar rotation rates and appear to persist for many months in the region.

One such eddy, seeded with drifters in June 1998, was observed to persist for at least seven months (**Figure 4**). It initially drifted slowly southwestward but later returned northwestward towards the outer islands of the archipelago. Other large anticyclones have been observed downstream of the island pair, Tenerife and La Gomera, trapped close to the islands for at least several weeks before moving away from the island. Smaller cyclones and anticyclones are frequently seen in sea surface temperature images being spun off from the flanks of the other smaller islands.

The generation of these eddies may not be entirely a result of the oceanic flow past the islands. The Canaries are high volcanic islands situated in a regime of strong southwestward trade winds. The high island peaks block the flow of the trade winds to form extended lee regions downwind. These are bounded by localized horizontal shear in the wind field and so are locations of strong Ekman divergence and convergence. Ekman transport takes place in the near surface layer and is to the right of the wind in the Northern Hemisphere. At the western boundary of the lee, upwelling of deeper waters must compensate the divergence, while at the eastern boundary, sinking must occur. The upwelling and downwelling elevates or depresses, respectively, the pycnocline from its unperturbed depth. Because the downwind scale of the lee is limited, the elevation or depression of the pycnocline tends to form an eddy of the same sign as expected from the current past the island. The vertical motions expected from the horizontal wind shear on the lee boundaries are of the order of tens of meters per day, and so potentially could contribute significantly to the observed eddy production.

Despite the frequent reports of eddies spun off from the islands, it has proven difficult to obtain time series observations of eddy generation which would allow determination of basic eddy properties such as average shedding frequency, size, propagation speed or whether eddies are shed alternately

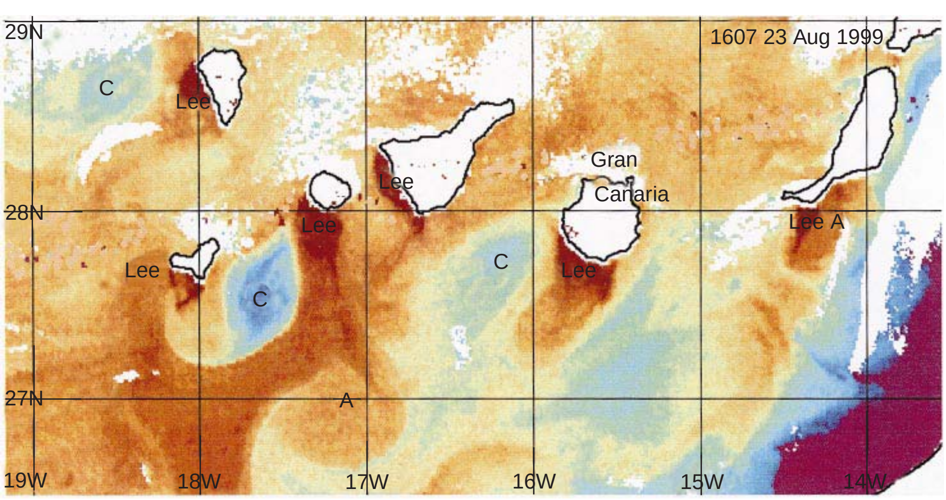

Figure 3 Sea surface temperature image showing multiple eddies shed from the Canary Islands during August 1999. Cyclones and anticyclones are labeled C and A, respectively.

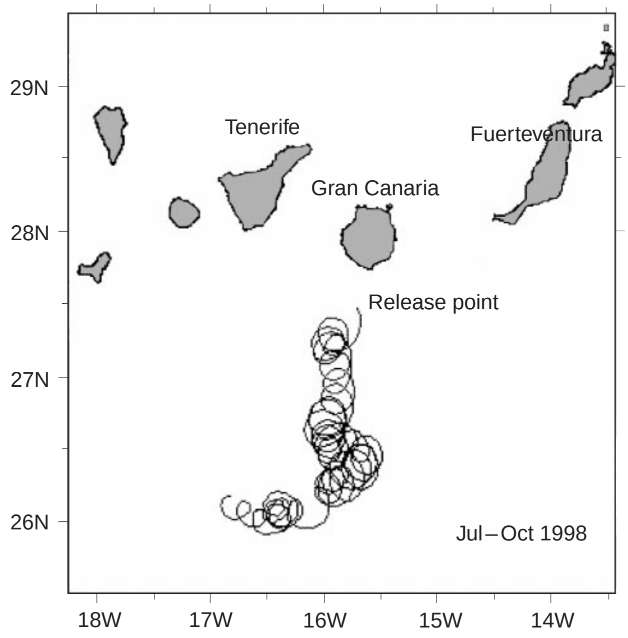

Figure 4 Path of a drifter released into an anticyclonic eddy shed from Gran Canaria in June 1998. The eddy persisted for 7 months though only the first three months are shown here (Courtesy of Dr Pablo Sangra, University of Las Palmas de Gran Canaria.)

from opposite island flanks. Remote sensing even in the subtropics is often blocked by cloud cover and time series *in situ* sampling is difficult to maintain.

A situation similar to the Canaries has been observed in the case of the Hawaiian archipelago, situated in the North Equatorial Current and trade winds of the Pacific. Downstream of these mountainous islands, the trade winds with speeds of 10–20 $m\,s^{-1}$ are separated from the calmer lee by strong boundaries of high wind shear. Locally, the depth of the surface mixed layer depends on wind speed: in the channels between islands, deep mixed layers are observed; in the lee, stirring by the wind is too weak to distribute solar heating down below the surface layer and intense surface warming results during the day. Sharp surface temperature fronts up to 4°C, are often associated with these wind shear lines, as is also observed in the Canaries. The lee of islands in both archipelagos is often visible in sea surface temperature images as a significantly warmer triangular area extending downwind.

Ekman transports associated with the wind pattern produce pycnocline perturbations as in the Canaries, resulting in intense anticlockwise eddies under the northern shear lines, and less intense clockwise eddies under southern shear lines. The depth of the mixed layer in the lee of Hawaii can vary from less than 20 m in the counterclockwise eddy to more than 120 m in the clockwise eddy. **Figure 5** shows a diagram of the Hawaiian island situation which applies equally well to the Canaries. Though the wind has long been viewed as an important generating mechanism for the Hawaiian eddies, it is still unclear how the variability of the wind field affects oceanic eddy generation. The wind itself has often been observed in satellite images of low level clouds to form wakes of counter-rotating atmospheric eddies behind Hawaii and the Canaries. The eddy-shedding period in this case is much shorter, about 10 h, than for oceanic eddies, which have a periodicity of many days. Presumably it is the wind field averaged over the timescale of the ocean eddies that is important. However, it is quite possible that atmospheric eddy shedding is intermittent;

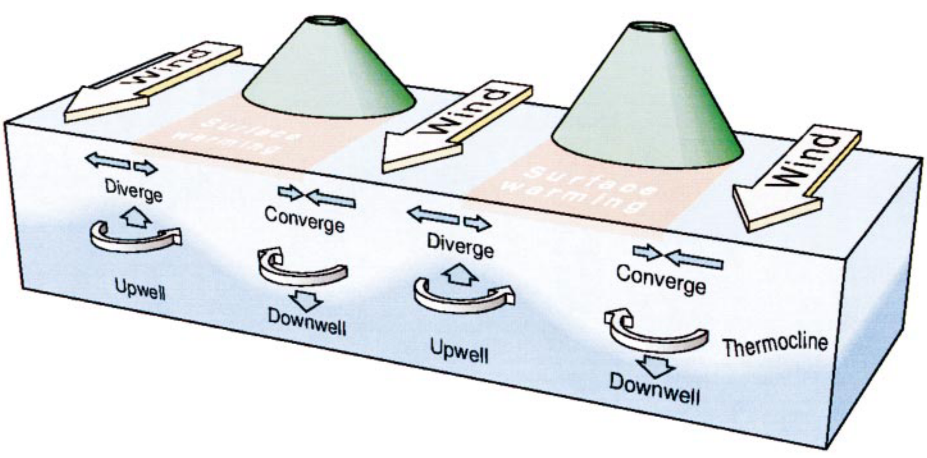

Figure 5 Hawaii schematic showing the mechanism of eddy generation by wind shear. Ekman surface layer transports lead to depression and elevation of the pycnocline in regions of convergence and divergence, respectively. This in turn generates oceanic anticyclones and cyclones. (Courtesy of Professor P. Flament, University of Hawaii.)

during periods when a trapped eddy regime dominates the wind shear lines are relatively stationary and able to feed energy into oceanic eddy production. Further observations are required to unravel the mechanisms at work in these island situations.

The North Equatorial Current impinging on the Hawaiian islands, of course, will also tend to produce eddies. The cumulative effect of the many eddies that are spun off is the formation of a mean large scale re-circulation behind the Hawaiian (cf. Barbados) chain that has been named the Hawaii Lee Counter Current. The longevity of individual eddies is further illustrated by one of their surface layer drifters which remained trapped in an anticyclonic vortex formed near the southern flank of the main island, drifting westward over 2000 km at $11\,cm\,s^{-1}$. This and other drifter tracks showed the remarkable phenomenon of vortex doubling, the process of merging of two vortices of the same sign. When two identical vortices merge, the radius of the merged eddy is $\sqrt{2}$ of the original and its period of rotation is doubled. The drifter appeared to show three such vortex-doubling episodes during its trajectory.

In the case of larger shallow sea islands in a tidal regime, the flow reverses before a wake can be properly set up. However, the relative motion between the island and surrounding body of water allows the island to act as a 'stirring rod' because of the increased flow speeds on the island flanks which produce vertical mixing within a tidal mixing front some distance off the shore. In a stratified region this introduces nutrients from the lower layers into the surface layers so enhancing productivity around the island. In the case of the Scilly Islands mixing between low density surface layer and high density bottom layer waters produced intermediate density enriched water spreading out between the layers causing enhancement of the mid-depth chlorophyll maximum in the area around. Similar results were found around St Kilda off western Scotland.

Conclusions

Island wakes produced by eddy shedding have been observed in both deep ocean and shallow shelf sea situations. In some cases of shelf sea islands there is no wake as such, because of weak mean flows, but vigorous tidal currents can produce regions of well-mixed water in the surrounding area. In the oceanic case, though there are many examples of eddy and wake observations, there has yet to be a definitive study demonstrating the phenomenon over a range of flow conditions.

The effect of islands on the flow regime does have biological consequences, in that enhanced mixing related to shallow sea islands can increase primary production. It has often been argued that island shed eddies may provide a mechanism for retaining fish larvae in the vicinity of their spawning areas. Although the evidence for this is not yet convincing the existence of mean return circulation has been demonstrated in both model and drifter studies of oceanic islands.

It was suggested that energy dissipation caused by island flow disturbance could account for 10% of the wind kinetic energy input to the Pacific. However, it is also considered that general turbulent processes known within the deep ocean may be too weak by more than an order of magnitude to explain global redistribution of energy input. In this case vertical mixing must be greater at the ocean boundaries with land than previously considered and the role of islands and island chains may be greater than is presently perceived. Simulations of the Barbados wake indicated that the flow disturbance was extensive, reaching at least eight island diameters downstream. Both the Canaries and Hawaii archipelagoes clearly control physical conditions for large distances downstream, modifying water masses through mixing, enhancing productivity and shedding long-lived eddies.

See also

Canary and Portugal Currents. Ekman Transport and Pumping. Fish Migration, Horizontal. Mesoscale Eddies. Pacific Ocean Equatorial Currents. Upper Ocean Vertical Structure. Wind Driven Circulation.

Further Reading

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ISOTOPES

See **COSMOGENIC ISOTOPES. NITROGEN ISOTOPES IN THE OCEAN. OXYGEN ISOTOPES IN THE OCEAN. PLATINUM GROUP ELEMENTS AND THEIR ISOTOPES IN THE OCEAN. RARE EARTH ELEMENTS AND THEIR ISOTOPES IN THE OCEAN. URANIUM–THORIUM SERIES ISOTOPES IN OCEAN PROFILES.**