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Johnston Atoll's Wake[1]

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ABSTRACT

In February 1968, when the incident flow was 60 cm/sec toward the northwest, a wake resembling a von Kármán vortex street was present in the ocean downstream of Johnston Atoll (16°45′N, 169°31′W). In October 1968, when the incident flow was 15 to 20 cm/sec, no wake was detected. The February wake system moved downstream as a unit at 45 cm/sec, with a period of 4 days, a wavelength of 160 km, and a width between vortex rows of 55 km. Two-dimensional wake theory and a quantitative comparison with the strikingly similar atmospheric island wakes observed in satellite photographs suggest: that Johnston Atoll's vortex wake probably extended downstream some 600 km, or three wavelengths; that the kinetic energy was dissipated at a rate of 10^{21} ergs per eddy; and that this energy represented essentially all of the kinetic energy in the upstream flow over a width roughly twice the cross-stream diameter (26 km) of Johnston Atoll. To generate the energy dissipated by such a wake, a wind stress of 1 dyne/cm^2 would have to act continuously over an area of 2×10^4 km^2. Thus, whenever the incident flow exceeds the critical value for a vortex-wake formation, Johnston Atoll apparently dissipates wind-induced kinetic energy and disturbs the downstream flow over areas that are two orders of magnitude larger than its own area at the 100-fathom isobath. Effects of island wakes may therefore be very important within island groups and for some hundreds of kilometers downstream. However, all major island obstacles in the Pacific Ocean would dissipate less than 10% of that ocean's wind-induced kinetic energy even if all of them had vortex wakes at all times.

Introduction. A striking feature of the Pacific Ocean is the multitude of islands and island groups that dot its central and western reaches. Its islands alone would set that ocean apart from all others on this planet. How important are these island obstacles in modifying the flow of currents? Do they dissipate significant amounts of energy, thus playing in part the role often assigned to the western-boundary currents? Or are island effects confined to narrow boundary layers, with purely local importance?

The oceanographic literature has remarkably little to contribute regarding the above questions. Shtokman (1966) examined the effects to be expected, in

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theory, from island barriers that occupy various portions of a wind-driven gyre system. He did this in order to account for anomalous circulation that is often observed near islands. Uda and Ishino (1958) presented results of scale-model experiments that employed obstacles of various shapes to determine where upwelling, and consequent enrichment, might occur in the vicinity of islands. Neither Shtokman nor Uda raised the question of energy dissipation by islands.

Van Dorn, Hacker, and Lam (n.d.), who made current measurements near Swains Island and Manihiki Atoll, have suggested the presence of vortex wakes downstream of both islands, but their observations were not conclusive because of fluctuations in the incident flow and because of the short periods of observation.

No other field studies have been reported that could serve as tests of Shtokman's theoretical analysis or of Uda's scale-model experiments. In 1963, I considered accomplishing this, in part, by using historical data from the area near the Hawaiian Islands and by carrying out half a dozen cruises specifically designed to study the islands' boundary layer and wake. However, Hawaii's boundary layer and wake soon proved to be larger, more complex, and more variable than I expected; one or two ships cannot cover areas that are large enough synoptically to permit definitive analysis. Some of the eddies and countercurrents that characterize the circulation near the Hawaiian Islands may be accounted for by Shtokman's theory, others may be topographic systems of the kind discussed by Uda (Manar 1967), and still others are probably due to local horizontal wind shear generated by the venturi effect in inter-island channels (Patzert 1969).

Therefore, in 1968 I made a field study of a much simpler system: of the flow past a small isolated midocean barrier, Johnston Atoll (16°45'N, 169° 31'W). The results are presented here and are compared with island-wake phenomena on the same scale in the atmosphere.

The flow past Johnston Atoll, examined on two occasions, was timed to coincide with periods of strong and weak incident flow as reported by Kopenski and Wennekens (1966). In February and October 1968, the TOWNSEND CROMWELL and the CHARLES H. GILBERT[2] visited Johnston together. The CROMWELL was assigned the task of measuring large-scale flow while the GILBERT made current measurements near Johnston.

During the February cruise, a steady current, flowing northwest at about 60 cm/sec, moved past Johnston. Downstream, the flow strongly resembled a simple von Kármán vortex wake with a period of 4 days, a wavelength of 160 km, and a width between vortex rows of 55 km. This is believed to be the first von Kármán vortex wake detected in the ocean, although such wakes frequently have been observed in satellite photographs of cloud patterns (Hubert

2. Ships of the National Marine Fisheries Service, Southwest Fisheries Center, Honolulu Laboratory formerly the Bureau of Commercial Fisheries Biological Laboratory, Honolulu).

and Krueger 1962, Chopra and Hubert 1964, 1965, Lyons and Fujita 1968, and Stevenson and Nelson 1968).

Reports of von Kármán vortex wakes in the athosphere are no longer rare, but all are due to a special combination of circumstances: the wind, an obstacle, the right kind and quantity of cloud cover at the proper level, and a satellite overhead taking pictures; all must be present. Since atmospheric vortex wakes doubtless form under much less restrictive circumstances, they are apt to be fairly common. Their discovery merely awaited the use of the proper means of observation. Similarly, von Kármán wakes, like that observed downstream of Johnston, will doubtless prove to be commonplace when we begin to look for them in the ocean.

During the October cruise, incident flow was weaker, about 15 to 20 cm/sec, and no wake could be detected near Johnston.

Previous Studies near Johnston. Aside from the results of two intensive surveys carried out by Kopenski and Wennekens (1966), little is known about the currents near Johnston. The "U.S. Coast Pilot" mentions only a 1–2-knot wave-driven current flowing south in the ship channel. The "Atlas of Surface Currents, Northwestern Pacific Ocean" (U.S. Navy, Hydrographic Office 1950) presents a few scattered observations near Johnston: the chart for February shows drift toward the west or northwest at speeds of 10 to 17 nautical miles per day within a few degrees of Johnston; comparable values (with smaller northerly components) are shown for January and March; the October chart also shows west and northwest drift, but at speeds of only 3 to 12 nautical miles per day in the vicinity of Johnston; the September and November charts show similar direction and speed. A chart published by the Deutsche Seewarte (Anonymous 1942) shows moderately persistent currents flowing westnorthwest at speeds of less than 12 nautical miles per day near Johnston.

Kopenski and Wennekens (1966) measured flow over the reefs and the nearshore areas of the central and western portions of Johnston during January-February and July-August 1965. They concluded that the dominant factors that determined the flow were: tidal currents (east-west), wave-driven flow over the reef (toward the southwest), and the North Pacific Equatorial Current, which was fairly strong and steady (toward the west) in winter, but weak and variable in summer. Their observations were made close to shore, most probably within the island's boundary layer. Their conclusions therefore apply to events within the boundary layer and not to the offshore wake, if any. Kopenski and Wennekens gave no estimate of the speed of the net incident flow. From measurements made in the seaward portions of their survey, I deduce that the net flow in winter was at least 0.2 knot (current-meter data) and may have reached a speed as high as 0.9 knot (current crosses in deep water) toward the west or slightly north of west. In summer, no net flow was apparent.

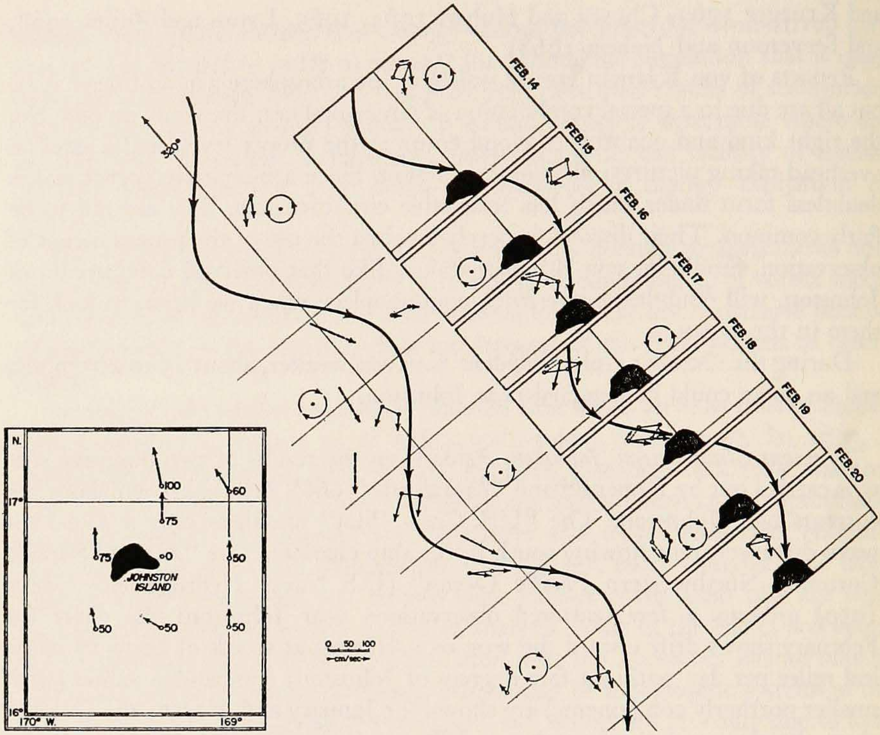

Figure 1. Currents near the sea surface off Johnston Atoll, February 1968. The individual panels in the upper right show observed longline drift during each day's fishing and the drift transformed to a frame of reference moving with the vortex wake (paired arrows). In the center, the transformed data have been arranged in sequence along a common axis. Single dashed arrows show the drift of the GILBERT as she lay to overnight. A von Kármán vortex street (sinusoidal line and circular arrows) has been fitted to the data. The inset in the lower left shows Johnston Atoll with 10- and 100-fathom depth contours and the drift (centimeters per second) of the CROMWELL during stations occupied on 12 and 13 February 1968.

Methods. In February, the CROMWELL measured the field of density for geostrophic computations, using an STD (salinity-temperature-depth recorder). At a few locations off Johnston it was possible to use radar to estimate the ships drift while on station, 12 and 13 February (Fig. 1, lower left). The GILBERT made direct measurements of near-surface currents based on the drift of longline fishing gear for one 8-hour period each day from 14 through 20 February (Fig. 1). Large drag in the water, minimal windage, and the possibility of detecting local current shear from differential motion of the two ends of the gear suggested the choice of longline as a current drogue.

In October, both ships made longline drift measurements and the CROMWELL ran a GEK (Geomagnetic Electrokinetograph) pattern (Fig. 2).

Only the longline drift data, the drifts of the two ships, and the GEK measurements provided usable information on Johnston Atoll's wake.

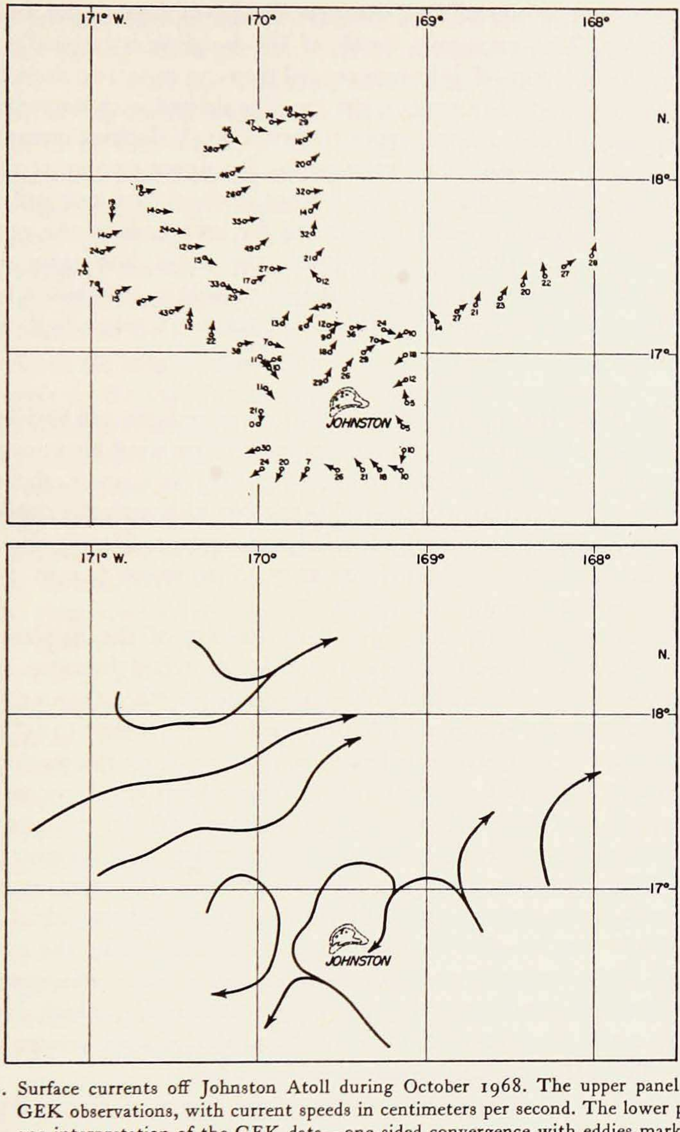

Figure 2. Surface currents off Johnston Atoll during October 1968. The upper panel shows the GEK observations, with current speeds in centimeters per second. The lower panel shows one interpretation of the GEK data—one-sided convergence with eddies marking the position of the front. Longline drift observations (not shown) agreed closely with the GEK data, but the drifts were usually too small (4 to 8 km) to be considered significant relative to the measurement errors.

For present purposes, the longline fishing gear (Yoshida 1966) is considered to be simply an 11-km length of 1-cm diameter line suspended from floats at intervals of 385 m. When set, the longline extended about 9 km from end to end. The estimated center of drag was near the 80-m depth—the mean depth of the mainline. The maximum depth of the longline is, typically, 200 m. Since mixed-layer depths off Johnston ranged from 70 m to 100 m in February and from 60 m to 80 m in October, the longline drifted with currents in both the surface mixed layer and the upper thermocline. Velocities measured with longline gear may be regarded as averages for the upper 150 m to 200 m.

At all stations, the longline gear was set out at dawn each day and retrieved eight hours later. Positions were determined for both ends of the gear at the time of setting, twice while the gear was in the water, and again when the longline was retrieved. Successive fixes agreed to within less than 2 km when Loran A alone was used and to within less than 1 km when both Loran A and radar were used.

Results. WINDS. During February, the winds were light and variable, never exceeding 7 m/sec (15 knots). The sea was calm. The wind blew mostly from the west or northwest at 5 m/sec (10 knots) or less. As a result, the effects of the wind on the longline are considered negligible, although the ship's drift as it lay to was clearly affected by both wind and current. During the October cruise, winds were from the northeast at 7 to 10 m/sec (14 to 20 knots). Seas were moderate to somewhat rough.

CURRENTS DURING FEBRUARY. The best estimate of the incident flow is provided by the drift of the CROMWELL as she approached Johnston (from the east) on 11 February and as she left Johnston (headed northeast) on 15 February. On both occasions she experienced drift toward the northwest (315°T) at 1.2 knots (60 cm sec^{-1}) with winds of less than 10 knots from the westnorthwest. The CROMWELL also drifted northwest at 60 cm sec^{-1} on the oceanographic station most distant from Johnston (Fig. 1, lower left). In the absence of a more definitive measurement, I have therefore settled on a velocity of 60 cm sec^{-1} toward the northwest as a reasonable estimate of the incident flow.

The panels in the upper right of Fig. 1 show (to scale) the initial and final positions of the longline gear for each day and the drift during the intervening eight hours. Since the wake moved downstream, these observations were transformed to a coordinate system that moved with the wake, which (see below) traveled 45 cm/sec (0.9 knot) toward 320°T. Pairs of solid arrows show the longline drifts in this moving system of coordinates.

These transformed data are also shown in the center of Fig. 1, arranged in the proper geographic and temporal sequence along a common axis that represents the path of Johnston Atoll through the water in this system of coordinates. In addition, the GILBERT's overnight drift as she lay to is shown (single broken arrows), similarly transformed and placed in the proper location for each night. A vortex street (sinusoidal lines and circular arrows) has been subjectively fitted to the data.

Finally, the insert at the lower left shows the geographic location of Johnston, the 10- and 100-fathom depth contours, and the drift (cm sec^{-1}) of the CROMWELL during oceanographic stations occupied on 12 and 13 February.

ANALYSIS OF FEBRUARY OBSERVATIONS. The speed and direction of the translation of the vortex wake were determined by trial and error: direction was taken to be that which yielded the best symmetry, with clockwise and counterclockwise eddies on opposite sides of the center line. Thereafter various speeds of translation were assumed; these ranged from 25 cm sec^{-1} (0.5 knot) to 75 cm sec^{-1}; only speeds of 40 to 50 cm sec^{-1} yielded plausible results. The best fit, using 45 cm sec^{-1}, is presented in Fig. 1.

Once the speed and direction of the wake had been estimated, a more critical analysis could be performed; the results are shown in Fig. 3. The upper panel presents the longline observations, as seen by an observer who is stationary with respect to the earth's surface, together with streamlines of a theoretical von Kármán vortex wake in the same frame of reference. The second panel shows the streamlines and the observations transformed to a coordinate system that moves 45 cm/sec^{-1} toward 320°T. The third panel shows longline drifts that would have resulted (if the flow had in fact been a pure von Kármán wake) from the two-dimensional theoretical equations (Lamb 1932: 224) applied at the midpoint of each day's drift. Finally, the fourth panel shows the residuals: those vectors which, when added to the theoretical drift, yield the observed values as resultants. Residuals have not been calculated for the GILBERT's drift because her shallow draft and large superstructure make her sail appreciably even in light winds.

CURRENTS DURING OCTOBER 1968. Fig. 2 shows the GEK measurements made from the CROMWELL during October. No wake is in evidence, probably because the incident flow is too weak—roughly 15 to 20 cm/sec (0.3 to 0.4 knot). Evidently Johnston was very close to a one-sided convergence, where a northwest flow approaching the island met a current flowing at right angles, toward the northeast. I interpret the eddies just west of, and around, Johnston to be part of a line of vortices associated with the frontal zone.

Discussion. JOHNSTON ATOLL'S WAKE. The simplest way to account for the observed flow during February 1968 is to assume that it was a von Kármán vortex-street wake moving toward 320°T at 45 cm/sec, with a period of 4 days, a wavelength of 160 km, and a width between vortex rows of 55 km. Residual drift values (Fig. 3, fourth panel) make it clear that this simple kinematic model accounts for most of the data.

Agreement between the model and the observations is remarkable, considering the fact that the model assumes constant strength, direction, and speed of translation of the wake system, two-dimensional flow, and no perturbations in the incident current.

The residuals for 19 and 20 February were large relative to the navigational error and are probably significant. However, on these two days, and only then,

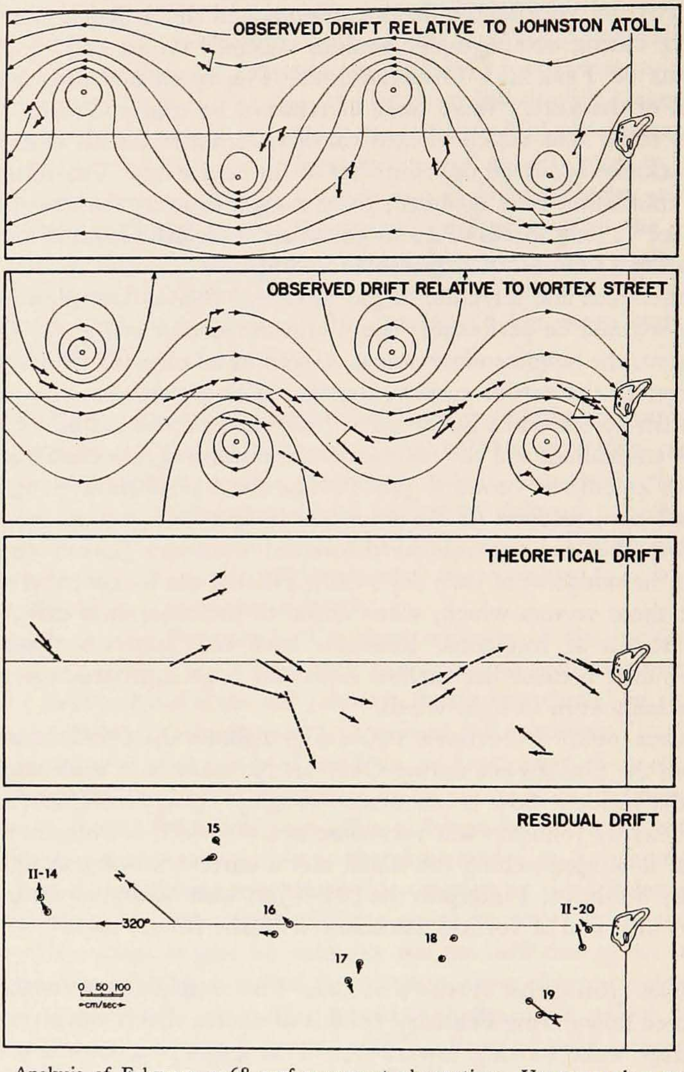

Figure 3. Analysis of February 1968 surface-current observations. Upper panel: a von Kármán vortex street moving toward 320°T, with observed longline drift data. Second panel: a stationary von Kármán vortex street fitted to the transformed longline drift data; the single broken arrows show the GILBERT's overnight drift in this same system of coordinates. Third panel: the theoretical drift at the same locations. Fourth panel: the residual drift, the difference between observed and theoretical values. Uncertainty due to navigational error is represented by small circles 4 km in diameter. Theoretical calculations are based on the values listed in Table 1.

did the longline happen to be set within less than 25 km of the center of an eddy. In the simple-point vortex model used to calculate residuals, the center of each eddy is a singular point having infinite velocity, and theoretical velocities near the center are certainly overestimates. A more complex and realistic model, such as the Rankine vortex (Rouse 1963), which has a core in solid rotation and a zero velocity at the center, might improve agreement between theory and observation. Patzert (1969) fitted a Rankine vortex to oceanic eddies near the Hawaiian Islands, but this requires more information than was obtained from the Johnston Atoll eddies.

Currents near Johnston in October 1968 were slower than those encountered in February. The Reynolds number in February was about 70 (Table I), so in October it must have been approximately 20, based on the relative speeds of observed flow on the two occasions and assuming that the eddy viscosity remained the same. In laboratory experiments, Thom (quoted in Goldstein 1965: chapter 13) found that a circular cylinder in a wide channel developed a vortex wake at Reynolds numbers slightly above 30. It is therefore not surprising that there was no vortex wake in October.

COMPARISON WITH ATMOSPHERIC WAKES. Lyons and Fujita (1968) analyzed a number of atmospheric wakes (clear areas in the stratus) downstream of islands in the Aleutian chain; they found that wakes did not form unless the obstacle had an effective diameter that was larger than 1 km, so that the flow had an eddy viscosity Reynolds number greater than about 7. Kiska, the largest island treated by Lyons and Fujita, had the highest eddy viscosity Reynolds number, 170, and was the only island with a von Kármán vortex wake. Amchitka and Agattu, the next largest, had eddy viscosity Reynolds numbers of 128 and 133, respectively, but they did not develop von Kármán vortex wakes.

Chopra and Hubert (1964, 1965) performed a detailed analysis of atmospheric von Kármán wakes downstream of the Grand Canary, Tenerife, and Madeira islands. Their analysis of Madeira's wake is of particular interest here because it is the most detailed quantitative treatment of such wakes to date. Wilkins (1968) extended their approach by deriving equations for kinetic energy losses; he also corrected a minor error in the Chopra and Hubert analysis.

The two-dimensional kinematic theory (Lamb 1932, Prandtl and Tietjens 1934, Goldstein 1965) gives not only excellent agreement with experimental results in the laboratory but also describes major features of the islands' wakes. Table 1 presents major parameters of the wakes downstream of Madeira and Johnston, calculated from the classical theory and from Wilkins' (1968) treatment of energy dissipation. A key element in the calculation of eddy viscosity is the use of Lin's dimensionless ratio, β. Lin (1954) found that recognizable vortex-street wakes occurred only when β ($\beta = S/R$, the ratio of the Strouhal number to the Reynolds number) falls within the range 1×10^{-3} to 2.5×10^{-3}.

The calculations in Table I are based on values for distances, velocities, and

Table I. Comparison of vortex street wakes in the ocean and the atmosphere.

Variable	Symbol	Johnston Atoll	Madeira	Equations used
Effective diameter of island	d	26 km	40 km	–
Wavelength or pitch	a	160 km	190 km	–
Width between eddy rows	h	55 km	83 km	–
Ratio h/a	h/a	0.34	0.43	–
Speed of incident flow	U_o	60 cm/sec	10 m/sec	–
Speed of translation of wake	U_e	45 cm/sec	7.5 m/sec	–
Period of eddy pair formation	T	96 hours	7.2 hours	–
Period half-pendulum day	–	42 hours	22 hours	–
Frequency of eddy pair formation	N	2.9×10^{-6} sec^{-1}	3.9×10^{-5} sec^{-1}	$T = N^{-1}$
Strouhal number	S	0.12	0.16	$S = Nd/U_o$
Reynolds number (eddy viscosity)	R	70	90	$R = U_o d/K$
Eddy viscosity	K	2.2×10^{6} cm^2/sec	4.4×10^{7} cm^2/sec	$K = \beta U_o^2/N$*
Drag coefficient	C_d	2.1	2.1	$C_d = h/d$
Circulation strength	k	6.1×10^{8} cm^2/sec	1.1×10^{10} cm^2/sec	$k = 2a(U_o - U_e)\coth\left(\frac{\pi h}{a}\right)$
Drag per unit height or depth (cm)	D	9.8×10^{9} g/sec^2	5.0×10^{9} g/sec^2	**
Vorticity at eddy center, t_o $(= T/2)$	ζ_o	1.3×10^{-4} sec^{-1}	16.6×10^{-4} sec^{-1}	$\zeta_o = k/4\pi Kt$
Planetary vorticity, $\varphi = 16°45', 32°30'$	f	0.42×10^{-4} sec^{-1}	0.78×10^{-4} sec^{-1}	$f = 2\omega\sin\varphi$
Ratio, ζ_o/f	–	3.1	21	–
Rate of energy dissipation per unit mass ($r = 20$ km, t_o)	ε	0.007 ergs/g/sec	34 ergs/g/sec	†
Dissipation per unit volume during one eddy lifetime	–	4,900 ergs/cm^3	840 ergs/cm^3	††
Decay e-folding time, $r = 20$ km	t_e	125 hours	6.3 hours	$t_e = r^2/4K$ [from #]
Age, oldest eddy	–	$3t_e$, = 375 hours	20 hours, $\simeq 3t_e$	–
Distance downstream, oldest eddy	–	~600 km, $\simeq$4a	540 km, $\simeq$3a	–

* The β referred to here is *not* df/dy, but a dimensionless ratio used by Lin (1954), which is an inherent property of the vortex street; it assumes a near-constant value (1.75×10^{-3} was used here) whenever a vortex wake is present.

** $D = C_d \dfrac{\varrho d U_o^2}{2} = \varrho h N k = \dfrac{\varrho k^2}{2\pi a} + \dfrac{\varrho K h}{a}(2U_e - U_o)$

† $\varepsilon = \dfrac{k^2}{8\pi^2 r^2 t}(1 - e^{-r^2/2Kt})$

†† $\varrho \displaystyle\int_{t_o}^{\infty} \varepsilon\, dt = \dfrac{\varrho k^2}{16\pi^2 K t_o}\left[1 + \sum_{1}^{\infty} \dfrac{\left(\dfrac{r^2}{-2Kt_o}\right)^n}{(n+1)(n+1)!}\right]$; $t_o = T/2$, $r = 20$ km

\# $\zeta = \zeta_o \exp\left(-\dfrac{r^2}{4Kt}\right)$.

periods of Madeira's wake as reported by Chopra and Hubert and of the wake observed downstream of Johnston.

The precision of numerical values in Table I must not be overestimated. Measurements of the wake geometry and velocity are crude, and the theoretical models are highly idealized. The values for energy dissipation, in particular, are no more than order-of-magnitude estimates.

The age of the oldest eddy downstream of Madeira (see Table I) proves to be nearly three times the *e*-folding time for decay at a radius of 20 km. Assuming that an oceanic wake of similar size will also last approximately three times the decay *e*-folding time for the same radius, I obtained an age of 375 hours for the oldest Johnston Atoll eddy.

The oldest eddy in Madeira's wake was 540 km downstream. This is equivalent to three wavelengths. In the case of Johnston, an eddy 375 hours old moving downstream at 45 cm/sec should travel a distance of about 600 km, or four wavelengths.

Table I reveals striking similarities between wake systems in two different fluids. There can be little doubt that fundamentally similar phenomena are involved in both cases. The similarity is the more striking because of differences in the observations on which they are based: data for Madeira's wake were obtained from satellite photographs whereas data for Johnston's wake were obtained from a one-week series of observations of drift. These represent the two fundamental approaches used in hydrodynamics: the Eulerian, based on the field of motion at a given instant of time, and the Lagrangian, based on trajectories observed over a period of time.

Madeira has a diameter that is 50% larger than Johnston Atoll (40 km versus 26 km), and the drag coefficients are the same for both islands. Accordingly, Madeira's wake is 50% wider. The wavelengths of the two wakes differ by only 20%; if their Strouhal numbers had been equal, the wavelengths would presumably have differed also by 50%. The Strouhal number is a function of the Reynolds number and of the geometry of the obstacle (Goldstein 1965).

Energy Dissipation. The amounts of energy dissipated by Johnston Atoll's wake in February 1968 are large enough to be of some significance. If it is assumed that each eddy had an effective radius of at least 40 km and that the flow, as measured by longline, extended to some 200-m depth, then each eddy contained a volume of 2.6×10^{18} cm^3 (2.6×10^{12} m^3) and dissipated a total of 1.3×10^{21} ergs during its lifetime—energy that it accumulated as it formed. Since an eddy formed every two days, the current performed work equal to at least 7.6×10^8 watts, or 1×10^6 horsepower when generating the wake.

The immediate source of this energy was the kinetic energy of incident flow, which ultimately was derived from the wind stress.

Incident flow during February 1968 moved at a speed of 60 cm/sec, so that its kinetic energy was 1800 ergs/cm^3 as compared with the energy dissipated in the wake, 4500 ergs/cm^3 (Table I), or 10^{21} ergs for each eddy. Assuming that the incident flow extended to a depth of 100 m, the energy dissipated downstream was equivalent to that contained in a 50-km-wide cross section of the upstream flow. Since Johnston Atoll perturbed the flow over a width of some 100 km, the wake dissipated half of the energy contained in the perturbed incident flow. Similarly, the wake dissipated essentially all of the energy contained in a cross-stream width that is equal to that of the distance between vortex rows (55 km), or twice the energy incident on Johnston Atoll (26 km in width).

To compare energy dissipated in the wake with energy added by wind stress, one could use the dynamic-energy equation (Sverdrup et al. 1942: 486), but it is simpler and more direct to use the drag-per-unit depth, essentially 1×10^{10} g/sec^2 per unit depth (Table I). If the wake eddies extended to a depth of 200 m, the total drag exerted on Johnston in February 1968 was 2×10^{14} dynes. To generate this force, a wind stress of 1 dyne/cm^2 would have to act over an area of 2×10^{14} cm^2, or 2×10^4 km^2. Johnston seamount has a horizontal cross-sectional area of <200 km^2 at 100 fathoms; in February 1968 its wake was dissipating energy acquired over an area that is two orders of magnitude larger. For the Pacific Ocean as a whole, with an area of 1.6×10^8 km^2 (Sverdrup et al. 1942), 8000 vortex wakes the size of Johnston Atoll's would dissipate all of the energy introduced by wind if the mean wind stress was 1 dyne/cm^2.

A preliminary count of midocean island obstacles in the Pacific shows that they number no more than a few hundred. Thus, even if each island had a vortex wake at all times—which Johnston Atoll, at least, does not have—less than 10% of the Pacific Ocean's wind-induced energy would be dissipated. Clearly island wakes are not a major factor in the ocean's energy balance.

Nevertheless, island wakes may have large-scale effects wherever clusters of island obstacles are spaced less than two diameters apart. In such cases, the group could dissipate virtually all incident kinetic energy whenever the eddy viscosity Reynolds number exceeds the critical value for vortex-wake formation. Examples of such island groups are the Tuamoto Archipelago, the Fiji group, the New Hebrides and Solomon islands, the Gilbert and Marshall groups, and possibly the Caroline, Mariana, and Ryukyu groups, and the high islands of Hawaii. Water probably flows freely through such island groups until some critical velocity is reached; at higher velocities, most of the flow may well be deflected entirely around the group. Such a marked qualitative change in response to a minor quantitative increase in velocity would have widespread consequences, influencing not only the potential and kinetic energy but also the distributions of physical, chemical, and biological properties within the island group and for many hundreds of kilometers downstream. For example, the dispersal of pelagic larvae spawned by coastal animals would be strongly affected by the altered pattern of advection, as would sedimentation. Changes in advection would also modify the effects of air-sea heat exchange on the surface layers.

Conclusions. Evidence presented here strongly suggests that, in February 1968, Johnston Atoll had a von Kármán vortex wake. This wake had a period of 4 days, a wavelength of 160 km, and a width of 55 km between vortex rows. The Atoll's drag coefficient was 2.1, the Strouhal number 0.12.

The quantitative consequences of this hypothesis are presented in Table I, where a two-dimensional kinematic wake theory was used to estimate such parameters as the Reynolds number, eddy viscosity, circulation strength, drag-vorticity and energy dissipation, as well as the downstream duration of the wake: 600 km, or four wavelengths.

Table I also demonstrates that Johnston Atoll's vortex wake was markedly similar to the vortex wakes revealed by cloud patterns downwind of islands, as seen in satellite photographs.

In October 1968, no appreciable wake was present, presumably because the incident flow was only 15 to 20 cm/sec compared with 60 cm/sec in February; as a result the Reynolds number (based on an eddy viscosity) for October was approximately 20 compared with a value of about 70 for February.

By analogy with laboratory wake studies, one can conclude that there is a critical Reynolds number for Johnston Atoll that is greater than 20 but less than 70; here a transition occurs from a small-scale weakly turbulent wake regime to another regime in which large vortices are shed periodically, generating a von Kármán wake.

Johnston Atoll's vortex wake dissipated kinetic energy at an estimated rate of 10^{21} ergs per eddy. The energy thus dissipated was approximately equal to that contained in a 50-km-wide band of the upstream flow, or twice the cross-stream width of Johnston Atoll. To generate such energy, a wind stress of 1 dyne/cm^2 would have to act over an area of 2×10^4 km^2.

For the Pacific Ocean as a whole, energy dissipated by islands doubtless amounts to less than 10% of that generated by wind stress. However, island wakes and their influence on the distributions of properties are probably very significant in and near closely spaced island groups, such as the Marshall Islands, the Tuamotu Archipelago, and perhaps for the entire western tropical Pacific Ocean as well.

If island wakes are in fact analogous to wakes observed in the laboratory, then minor changes in the ocean-current velocity can cause major changes in the patterns of flow and in the energy dissipation near islands as the wake undergoes transition from a "laminar-flow" condition to large-scale periodic vortex shedding. It is therefore important that we learn more about island-wake phenomena, in particular: (1) determine the effects of planetary vorticity, which could influence the trajectory of the wake's eddies (Warren 1967) and perhaps suppress formation of vortices of one sign until the local value of the vertical component of planetary vorticity is exceeded (Pohle et al. 1965), thus making the critical Reynolds number a function of latitude;[3] (2) ascertain how wakes generated by groups of islands arranged in a line (Hawaiian Islands, Mariana Islands) or in clusters (Marshall Islands, Tuamotu Archipelago) differ from those generated by isolated island obstacles; and (3) determine whether the kinetic energy in the wake is carried down the spectrum of turbulence until it is lost as heat (as would be the case if the ocean were a homogenous medium) or whether appreciable amounts of kinetic energy are converted to potential energy by divergence and vertical mixing.

Finally, the presence of orderly intermediate-scale vortex wakes in the lee of islands raises a fundamental question in fluid dynamics: Why should the flow of stratified rotating fluids at Reynolds numbers of the order of 10^{10} or higher so strikingly resemble the flow of homogeneous nonrotating fluids at Reynolds numbers of the order of 10^2? The use of an eddy viscosity coefficient to resolve this difficulty by producing reasonable numerical values is an artifice that sheds no light on the fundamental problem: concentrated vorticity generated by the flow of a stratified rotating fluid past an internal boundary. The Rossby number is relatively high (about 0.3 for both Madeira and Johnston Atoll), so the problem is inherently nonlinear.

White (1971a, 1971b; see also McKee 1971) has dealt with barotropic flow past a circular obstacle on a β plane, which, for eastward flow, gives rise to a standing Rossby wake downstream. However, White's solution is not applicable to the Johnston Atoll or Madeira wakes, where the flow has westward components (easterly, to meteorologists). In such cases, White's "Island number", $Is = (\beta d^2/4U_0)^{1/2}$, is imaginary ($U_0$ is negative) and no standing Rossby wake can form. Even if the flow were toward the east, both Johnston Atoll and Madeira are too small to generate Rossby wakes at the velocities that cause vortex-wake formation. According to White (1971a), appreciable beta effects are to be expected only when the "Island number" exceeds unity. For Johnston Atoll and Madeira, the "Island number" can be expected to have magnitudes of only 10^{-1} and 10^{-3}, respectively. White's inertial model does not account for the wakes discussed here.

Evidently vortex wakes on an intermediate scale in nature, like those observed in the laboratory, are boundary-layer phenomena involving friction as well as inertia. Theoretical work on island wakes must therefore consider not only the Rossby number ($U/\Omega L$) but also the Ekman number ($\nu/\Omega L^2$). Note that the ratio of these dimensionless numbers is the Reynolds number (UL/ν), an important parameter of wake systems both large and small. Wakes in nature probably resemble those on a much smaller scale in the laboratory

3. Observations from the Aleutian Islands by Lyons and Fujita (1968) yield a critical Reynolds number between 133 and 170 (52°N) as compared with a value between 20 and 70 for Johnston Atoll (about 17°N).

because dissipation of energy by friction is similar, in some significant sense, on both scales. Thus the island-wake problem reduces to the well-known problem of the nature of large-scale energy dissipation in turbulent fluids. Island wakes may prove to be ideal for the study of this important subject.

REFERENCES

ANONYMOUS
1942. Weltkarte zur Übersicht der Meeresströmungen. No. 2802. Dtsch. Seewarte, Berlin; 1 p.

CHOPRA, K. P., and L. F. HUBERT
1964. Kármán vortex-streets in earth's atmosphere. Nature, London, *203*: 1341–1343.
1965. Mesoscale eddies in wake of islands. J. atmos. Sci., *22*: 652–657.

GOLDSTEIN, SIDNEY (Ed.)
1965. Modern developments in fluid dynamics. Dover Publ., New York, Vol. 2; 702 pp.

HUBERT, L. F., and A. F. KRUEGER
1962. Satellite pictures of mesoscale eddies. Mon. Weath. Rev., Wash., *90*: 457–463.

KOPENSKI, R. P., and M. P. WENNEKENS
1966. Circulation patterns Johnston Island winter–summer 1965. U.S. Nav. oceanogr. Office, Washington, D.C., SP-93; 240 pp.

LAMB, HORACE
1932. Hydrodynamics. Dover Publ., New York. 738 pp.

LIN, C. C.
1954. On periodically oscillating wakes in the Øseen approximation, *In*, Studies in mathematics and mechanics presented to Richard von Mises, pp. 170–176: Academic Press, New York. 353 pp.

LYONS, W. A., and TETSUYA FUJITA
1968. Mesoscale motions in oceanic stratus as revealed by satellite data. Mon. Weath. Rev. Wash., *96*: 304–314.

MANAR, T. A.
1967. Progress in 1965–66 at the Bureau of Commercial Fisheries Biological Laboratory, Honolulu. Circ. Fish Wildl. Serv., Wash., 274; 51 pp.

MCKEE, W. B.
1971. Comments on "A Rossby wake due to an island in an eastward current." J. phys. Oceanogr., *1*: 287.

PATZERT, W. C.
1969. Eddies in Hawaiian waters. Hawaii Inst. Geophys. Rep. 69–8; 51 pp., 71 fig., 1 table.

POHLE, J. F., A. K. BLACKADAR, and H. A. PANOFSKY
1965. Characteristics of quasi-horizontal mesoscale eddies. J. atmos. Sci., *22*: 219–221.

PRANDTL, LUDWIG, and O. G. TIETJENS
1934. Applied hydro- and aeromechanics. Dover Publ., New York. 311 pp.

ROUSE, HUNTER
1963. On the role of eddies in fluid motion. Amer. Sci., *51*: 284–314.

SHTOKMAN, V. B.
1966. A qualitative analysis of the causes of the anomalous circulation around oceanic islands. Izvestiya, Atmospheric and Oceanic Physics, *2*: 1175–1185. [Translated by Allen B. Kaufman U.D.C. 551.465.52, p. 723–728.]

STEVENSON, R. E., and R. M. NELSON
1968. An index of ocean features photographed from Gemini spacecraft. Natl. Aeronaut. Space Admin., Manned Spacecraft Center, Houston, Texas, 348 pp. [Contr. 253, Bureau of Commercial Fisheries Biological Laboratory, Galveston, Texas.]

SVERDRUP, H. U., M. W. JOHNSON, and R. H. FLEMING
1942. The oceans; their physics, chemistry, and general biology. Prentice-Hall, Englewood Cliffs. 1087 pp.

UDA, MICHITAKA, and MAKOTO ISHINO
1958. Enrichment pattern resulting from eddy systems in relation to fishing grounds. J. Tokyo Univ. Fish., *44*: 105–129.

UNITED STATES NAVY. Hydrographic Office
1950. Atlas of surface currents, northwestern Pacific Ocean. H. O. Publ. 569, Washington, D.C., 12 pp. [Also H.O. Misc. 10,058-A.]

VAN DORN, W. G., P. W. HACKER, and R. K. LAM
n.d. Circulation around oceanic islands. Summary Report, Contract NSF GA804, SIO Ref. Ser. 67–34, 14 pp. + 10 fig.

WARREN, B. A.
1967. Notes on translatory movement of rings of current with application to Gulf Stream eddies. Deep-sea Res., *14*: 505–524.

WHITE, WARREN
1971a. A Rossby wake due to an island in an eastward current. J. phys. Oceanogr., *1*: 161–168.

1971b. Reply. J. phys. Oceanogr., *1*: 288–289.

WILKINS, E. M.
1968. Energy dissipated by atmospheric eddies in the wake of islands. J. geophys. Res., *73*: 1877–1881.

YOSHIDA, H. O.
1966. Tuna fishing vessels, gear, and techniques in the Pacific Ocean, *In*, Thomas A. Manar (Ed.), Proceedings, Governor's Conference on Central Pacific Fishery Resources, State of Hawaii; pp. 67–89.