

On the Surface Circulation in Some Channels of the Canadian Arctic Archipelago.

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ABSTRACT. This paper advances an explanation for the presence of surface currents in opposite directions on facing sides of some of the main channels of the Canadian Arctic Archipelago. It is found that geostrophic dynamics coupled with geometrical constraints and the general direction of surface drift through the archipelago can readily account for the existence, if not all the properties, of the observed flow patterns.

INTRODUCTION

The large scale surface flow through the Canadian archipelago is generally considered to be from the Arctic Ocean towards the south and east. Recent reviews by Herlinveaux (1974) and Walker (1977) agree in principle with the map compiled from historical data by Collin (1963), and reproduced here as

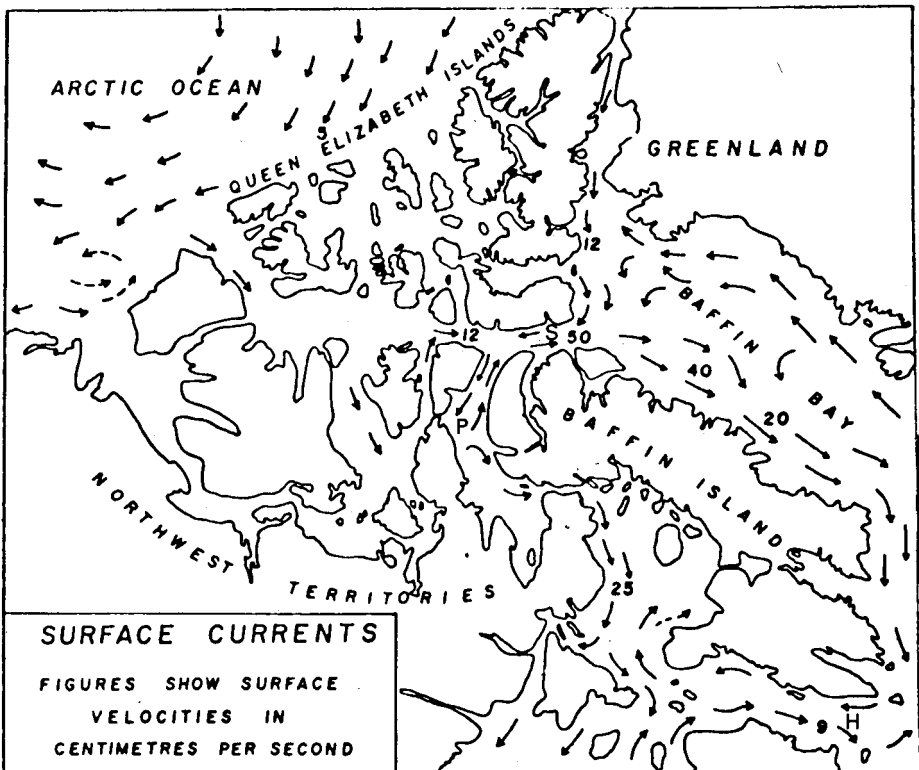


FIG. 1. Surface flows in the Canadian Arctic archipelago, as summarized by Collins (1963). The letters indicate the positions of the following sea-straits: H = Hudson Strait; S = Lancaster Sound; P = Prince Regent Inlet.

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Figure 1. Although this map shows the general south-eastward trend mentioned above, closer scrutiny reveals a number of counter-currents: in some channels (Hudson Strait, Lancaster Sound, Prince Regent Inlet), a flow on one side in the direction of the general trend is opposed by a current in the reverse direction on the other side of the channel. Similar re-entrant flows have been documented more fully in recent studies with moored current meters and satellite-tracked drifting buoys near the mouths of Hudson Strait (Osborn *et al.*, 1978) and Lancaster Sound (Fissel and Marko, 1978); an example is shown in Figure 2. In spite of differences in the details of the geometry of the various locations where the re-entrant circulations are observed, a profound dynamic similarity must be recognized between the different instances of the kind of flow pattern seen in Figure 1 and Figure 2. An understanding of the conditions responsible for the observed circulations facilitates the interpretation of data and extrapolation to as yet unsurveyed areas.

This paper presents a qualitative explanation of steady flow patterns at the junction of sea-straits in terms of geostrophic dynamics. It will be shown that the observed circulations may be attributed to the relatively large width of many of the channels of the Arctic archipelago with respect to the internal Rossby radius of deformation.

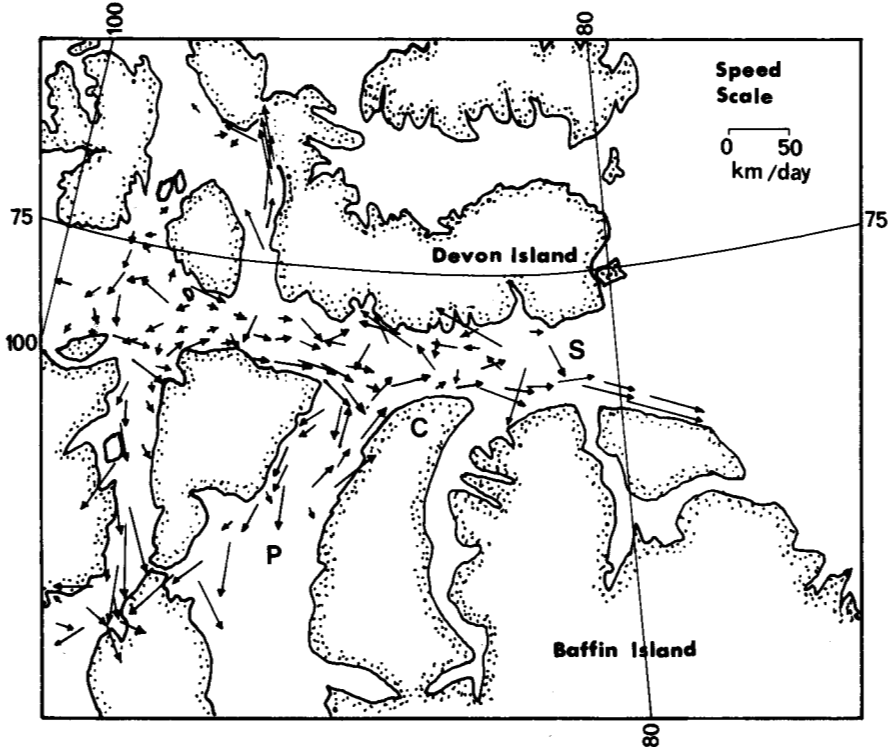


FIG. 2. Surface currents, inferred from the tracks of drifting buoys, as measured by Fissel and Marko (1978), in the Lancaster Sound area. Note the presence of current reversals on opposite shores near the mouth of Prince Regent Inlet (P) and in Lancaster Sound (S) north of the Brodeur Peninsula (C).

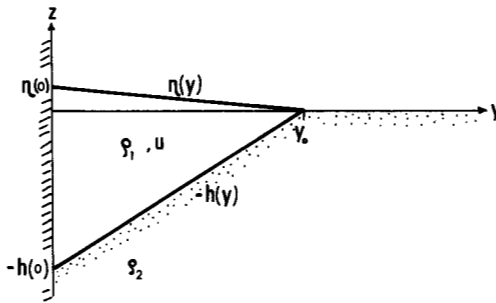


FIG. 3. Schematic cross-section of a coastal upper layer flow of speed u (out of the page) driven by a sea-surface slope $\eta(y)$; the lower layer, of density ρ_2 is at rest because the interface $h(y)$ slopes in a direction opposite to that of the free surface. y_0 is the distance from the coast at which the thickness of the upper layer vanishes.

Coastal Current Model

Consider a stratified flow in a channel of width L and total depth H . The current is assumed to be limited to an upper layer of uniform density ρ_1 overlying a lower layer of density ρ_2 which remains at rest (Fig. 3). The steady uniform flow in the upper layer is in geostrophic balance, with the speed u related to the surface slope through

$$u = \frac{g}{f} \frac{\partial \eta}{\partial y} \quad (1)$$

where g is the acceleration of gravity, f the Coriolis parameter and η the free surface displacement from a geopotential level. The condition of no-flow in the lower layer requires that the pressure gradient vanish below the interface $z = -h(y)$; under hydrostatic conditions, the interface must then have a slope given by

$$\frac{\partial h}{\partial y} = \frac{-\rho_2}{(\rho_2 - \rho_1)} \frac{\partial \eta}{\partial y} \quad (2)$$

The upper layer thus thins out in the cross-channel direction, since the interfacial slope is opposite to that of the free surface slope. If the channel is wide enough, the interface will intercept the free surface at some value of y less than L and the current will be limited to a wedge on one side of the channel (the right hand side, looking downstream).

Let the thickness of the upper layer be denoted by $t = h + \eta$; denoting the thickness at the wall ($y = 0$) by $t(0)$, where $t(0) = h(0) + \eta(0)$, we find from (1) and (2), that

$$t(y) = t(0) - \frac{fu}{g} \frac{\rho_1}{(\rho_2 - \rho_1)} y \quad (3)$$

The distance y_0 from the wall at which $t = 0$ is thus

$$y_0 = \frac{gt(0)}{fu} \frac{(\rho_2 - \rho_1)}{\rho_1} \quad (4)$$

On the other hand, the speed of interfacial waves in a fluid of the same density contrast is given by (LeBlond and Mysak, 1978, p. 76)

$$c^2 = \frac{g(\rho_2 - \rho_1)t(0)[H - t(0)]}{\rho_2 t(0) + \rho_1 [H - t(0)]}, \quad (5)$$

which, for an upper layer which is relatively thin with respect to the lower layer ($t(0) \ll H$) reduces to

$$c^2 = \frac{g(\rho_2 - \rho_1)t(0)}{\rho_1}. \quad (6)$$

Introducing the internal Rossby radius of deformation R and the internal Froude number F through

$$R = c/f; \quad F = u/c; \quad (7)$$

we find that

$$y_0 = R/F. \quad (8)$$

For a current system which is not completely restricted to the upper layer, the compensating slope of the interface will not be as steep and the flow wedge will be wider. Departures from an idealized two-layer system will have a similar broadening effect, so that y_0 is a lower bound on the width of a coastal geostrophic current. In the extreme case where lateral mixing determines the width of the density structure and of the associated geostrophic flow, the lateral scale is also given by the Rossby radius (Allen, 1973).

Application to the Arctic Archipelago

Let us now turn to Arctic data and compare the width of various current systems, as inferred from the local stratification, to the width of the channels where they occur. As a first example, consider the density field in a transverse section near the mouth Hudson Strait, as shown in Figure 4. We idealize the density structure on the south side of that strait in terms of an upper layer density $\rho_1 = 1026.0 \text{ kg/m}^3$ and a lower layer with $\rho_2 = 1026.4 \text{ kg/m}^3$; with $t(0) = 175 \text{ m}$ and $u = 0.4 \text{ m/sec}$ (from Osborn *et al.*, 1978), we find $y_0 = 13.5 \text{ km}$, which is clearly an underestimate of the distance from the coast at which the $\sigma_t = 26.4$ isopycnal surfaces in Figure 4. A similar calculation on the north side of Hudson Strait, with $(\rho_2 - \rho_1)/\rho_1 = 2 \times 10^{-4}$, $t(0) = 75 \text{ m}$ and $u = 0.25 \text{ m/sec}$ gives $y_0 = 4.7 \text{ km}$. This is again seen to be an underestimate, but less grossly so than on the southern side of the strait.

The width of geostrophic currents near the mouth of Lancaster Sound may be inferred from the dynamic topography (at the surface with respect to 500 decibars) presented by Muench (1971) and reproduced here as Figure 5. Although much variability is in evidence, the geostrophic flow pattern of 1961 shows some of the characteristics of the flows discussed here, with a coastal current westward near the north coast and an eastward current on the south coast, the two being partly connected across the sound.

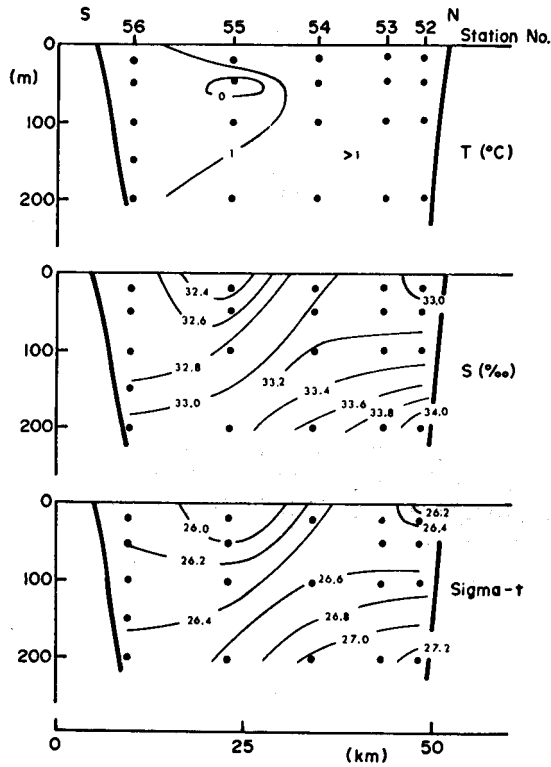


FIG. 4. Temperature, salinity and density sections across the mouth of Hudson Strait, Sept. 1977 (From Osborn *et al.*, 1978).

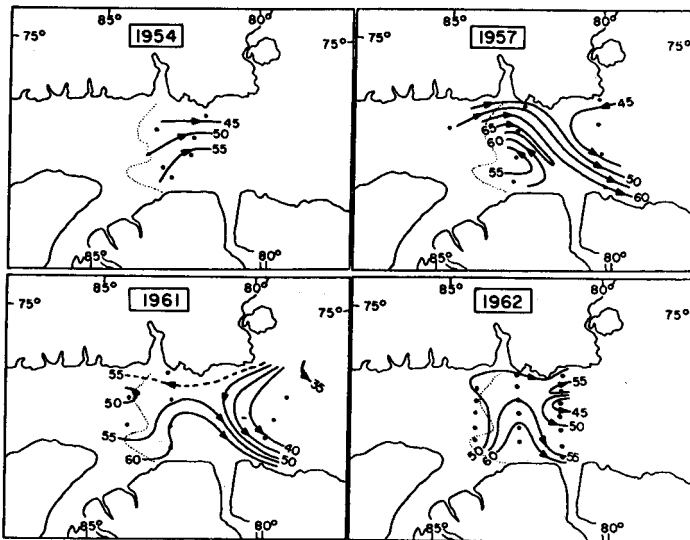


FIG. 5. Dynamic topography of the surface relative to 500 decibars in Lancaster Sound for various years, with a contour interval of 5 dyn. cm. The dotted line is the 650 isobath. (From Muench, 1971).

It does appear that across some of the main channels of the Arctic Archipelago there is enough room for two geostrophically balanced upper layer flows. This result does not of course explain why the observed currents are present: it merely shows that they can coexist without interfering with each other. Let us now consider the cause of their presence.

A purely geostrophic current flowing along a smooth but not necessarily rectilinear coast conforms to the shape of the coast. The coast is a streamline of the flow, since no water can go through it; pressure is uniform along streamlines in a geostrophic flow, so that the pressure (and hence the mean sea-level) is uniform along the coast. The surface slope which supports the current remains normal to the coast as the streamlines curve to follow the latter (Fig. 6a). As long as the radius of curvature r of the coast is large enough that the Rossby number u/fr remains well below unity, inertial accelerations are negligible and a geostrophic flow can turn corners (as in Fig. 6b) without separating from the coast. A side-channel of width $L \gg y_0$ appears infinitely wide and is penetrated by a coastal current irrespective of the presence of another coastal current flowing out of that channel (as in Fig. 6c). The general southeasterly outflow through the Arctic Archipelago brings coastal currents past the mouths of wide channels with (as in Lancaster Sound and Hudson Strait) and without (Prince Regent Inlet) significant coastal currents flowing out of them. The presence of reverse flows in these channels is thus to be interpreted as a consequence of the influence of the local geometry on the basic geostrophic dynamics of the large scale current system.

One should be careful at this point not to jump to unwarranted conclusions on the universality and ubiquity of the flow patterns discussed above. First of all, surface currents are affected by the wind, and upper-layer responses to wind forcing can undoubtedly often mask any underlying geostrophic flow pattern. Secondly, a coastal geostrophic flow such as the one illustrated in Figure 3 is subject to barotropic and baroclinic instability mechanisms

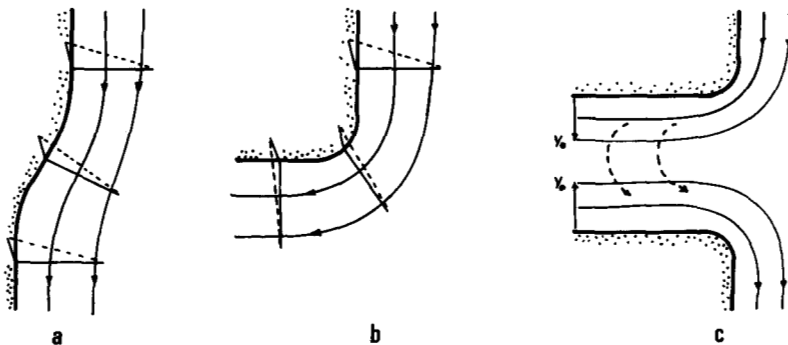


FIG. 6. Plan view of an upper layer geostrophic flow (as seen in cross-section in Fig. 3) flowing along a non-rectilinear coast (a) and around a corner (b). Dashed lines show the sloping free surface across the current. The penetration of a geostrophic flow into a side channel from which also issues a coastal flow is shown in (c); the two currents do not interfere with each other if the width of the channel exceeds $2y_0$, as shown. The dotted arrows show the cross-channel flow which is often observed in such situations (see Fig. 7).

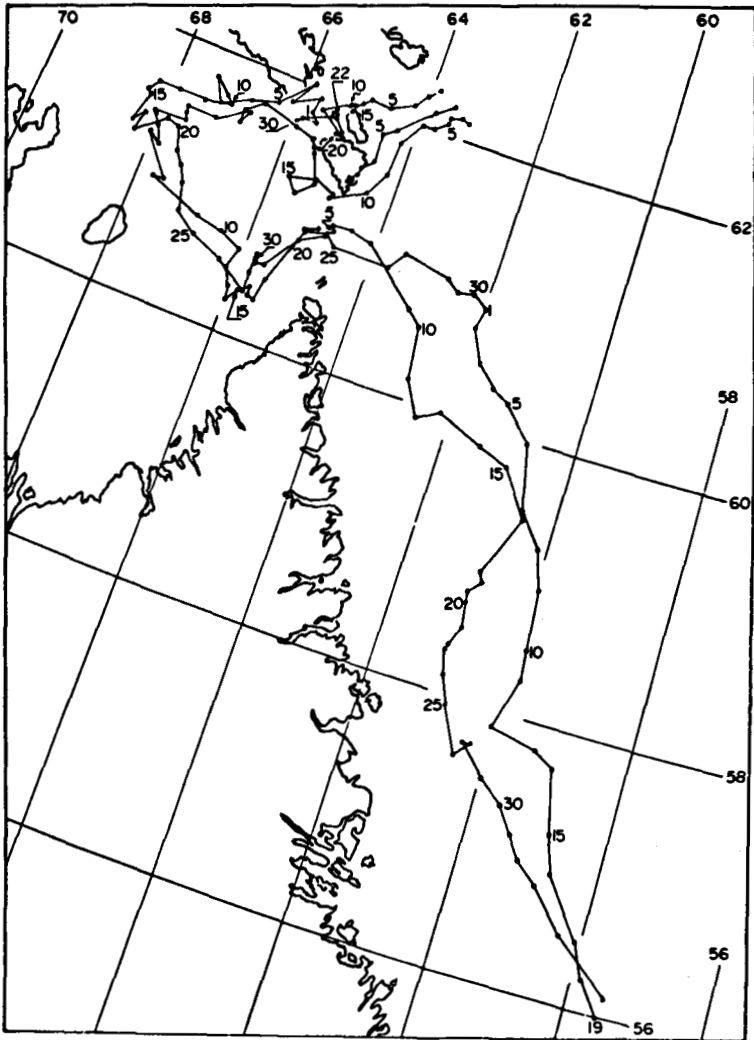


FIG. 7. Tracks of three surface drogues launched north of the mouth of Hudson Strait, near 62°N , 64°W on 2 Aug. 1977. Numbers denote dates. (From Osborn *et al.*, 1978).

(LeBlond and Mysak, 1978, Ch. 44) which will cause it to meander and to detach itself from the coast. Nonlinear effects associated with the curvature of the flow around corners can also lead to overshooting and occasional eddy shedding. Although the grid of stations from which Muench's dynamic topographies of Figure 5 are drawn is rather coarse for assessing the details of features of scales comparable to the Rossby radius, we may take the meandering visible in some of the panels of Figure 5 as evidence of the variability to which a coastal geostrophic current is susceptible.

Furthermore, it appears that even in those channels such as Hudson Strait where the reverse flow is best documented, there is a clear inter-relation

between the two coastal geostrophic currents. As is strikingly evident in the drogue tracks of Figure 7 (these are not isolated instances: see Osborn *et al.*, 1978, for additional examples), a cross-channel drift carries surface waters from the incoming northern current to the outflowing southern current. The causes of the cross-channel flow are not investigated here; it may be due to non-geostrophic effects such as that of friction (which causes flow down the pressure gradient) or of inertia (which leads to overshooting in corners). The transverse surface flow is responsible for the counter-clockwise aspect of the estuarine outflow of the strait near its mouth.

The broad framework of geostrophic flow adjustment to changing channel geometry developed above is thus to be considered as a lowest order explanation for some of the circulation patterns seen in the Canadian Arctic Archipelago; it has certainly no pretension of explaining all the vagaries of the surface flow patterns observed in those waters. It is also of interest to note that the Canadian Arctic Archipelago is probably the only region in the world endowed with a multiplicity of channels of width sufficiently larger than the internal Rossby radius for the type of circulation described to manifest itself. The East Indian Archipelago is the only other extensive area of comparable configuration. For similar near-surface values of density stratification, the internal Rossby radius in Indonesian waters is an order of magnitude larger than in the Arctic Archipelago, since the Coriolis parameter f (which is proportional to the sine of the latitude) is so much smaller near the equator. Thus, although Indonesian waters are relatively lighter at the surface than surrounding Indian or Pacific Ocean waters (Gorshkov, 1974), and hence endowed with a certain degree of surface stratification, the inter-island channels in the equatorial archipelago will appear narrower in terms of the internal Rossby radius than their polar counterparts; the type of re-entrant flows observed in the relatively wide Arctic channels is thus less likely to occur in the relatively narrow equatorial passages.

Conclusions

It has been shown that some of the main channels of the Canadian Arctic Archipelago are wide enough to accommodate a pair of coastal upper-layer geostrophic flows. The presence of surface currents flowing in opposite directions on facing shores of some of these channels is then interpreted in terms of the mean surface flow through the archipelago, from the Arctic to the Atlantic Ocean. More specifically, the penetration of coastal currents, flowing in a direction opposite to that of the general surface transport through the archipelago, into wide channels such as Hudson Strait and Lancaster Sound is recognized as a consequence of the large width of these straits with respect to the local internal Rossby radius of deformation.

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