Simulation of the Wind-Forced Near-Surface Circulation in Knight Inlet: A Parameterization of the Roughness Length

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ABSTRACT

Month-long observations of along-channel velocity made close to the surface of Knight Inlet are used with a numerical model to estimate the roughness length z_0 on the water side of the air–sea interface. In analogy with a very common parameterization for z_0 on the air side of the air–sea interface, z_0 is parameterized in the numerical model as $z_0 = au_*^2/g$ where $u_* = (\tau/\rho)^{1/2}$ is the friction velocity, g is the acceleration due to gravity, τ is the wind stress, ρ is the density of water, and a is an empirical constant. It is found that $a \approx O(10^5)$ for the dataset from Knight Inlet, a value four orders of magnitude larger than the value commonly used to estimate z_0 on the air side of the air–sea interface. When compared to empirical estimates of the significant wave height H_s, it is found that $z_0 \approx O(H_s)$. Further evidence is provided that a numerical model that uses the Mellor–Yamada level 2.5 turbulence closure scheme can simulate the near-surface, wind-forced circulation quite well.

1. Introduction

In the past few years, it has become evident that the oceanic boundary layer at the air–sea interface is significantly different than the atmospheric boundary layer on the other side of the air–sea interface (e.g., Craig and Banner 1994; Drennan et al. 1996; Terray et al. 1996; Craig 1996). One of the main reasons for this difference is that the breaking of surface waves causes enhanced levels of turbulent kinetic energy dissipation in the oceanic boundary layer. In fact, Terray et al. (1996) estimate that the breaking of surface waves enhances the energy flux to the oceanic boundary layer by an order of magnitude relative to what the flux would be for a boundary layer near a solid surface. They make the point that their results ‘‘demand new approaches to modeling the many processes of physical, chemical, and biological interest that are linked to the intensity of mixing in the very near surface layers.’’

Craig and Banner (1994) used the level 2.5 turbulence closure scheme of Mellor and Yamada (1982) with a simple model of the oceanic boundary layer to successfully reproduce the observations from a number of datasets. Their model predicts significantly enhanced surface turbulence relative to the standard ‘‘law of the wall.’’ They suggest that the roughness length z_0 in the oceanic boundary layer is of the order of the surface wave amplitude. This value for z_0 is much larger than what it is found to be on the atmospheric side of the air–sea interface. Craig (1996), using the same model, modified to simulate the laboratory experiment of Cheung and Street (1988), suggests that the z_0 increases with increasing wind speed and is of the order of one sixth the dominant surface wavelength. Craig (1996) notes that z_0 is a ‘‘significant scaling parameter for the dynamics in the wave-affected zone’’, but that ‘‘Dissipation measurements in the ocean or lakes are not yet precise enough to enable an accurate estimation of its magnitude.’’ For numerical modelers, it has long been known, for example, Blumberg and Mellor (1987), that the common assumption that the mixing length goes to zero at the surface, even when surface waves are known to be present, is not appropriate. Clearly, z_0 is an important length scale but it is as yet hard to quantify.

Stacey et al. (1995) developed a two-dimensional numerical model of Knight Inlet (a fjord located along the coast of British Columbia; Fig. 1) using the conventional 2.5 turbulence closure scheme of Mellor and Yamada (1982) and found that the model could successfully simulate much of the circulation in the inlet. The circulation in the fjord is significantly influenced by the winds, the tides, and freshwater runoff. Observations were made as shallow as 2 m from the surface, so the influence of the wind on the near-surface circulation was recorded, although the observations were not made rapidly enough to directly record the surface wave field. Stacey et al. found that, while the simulation was largely successful, the vertical shear in both the along-channel velocity and the density near the surface was too large to be phys-

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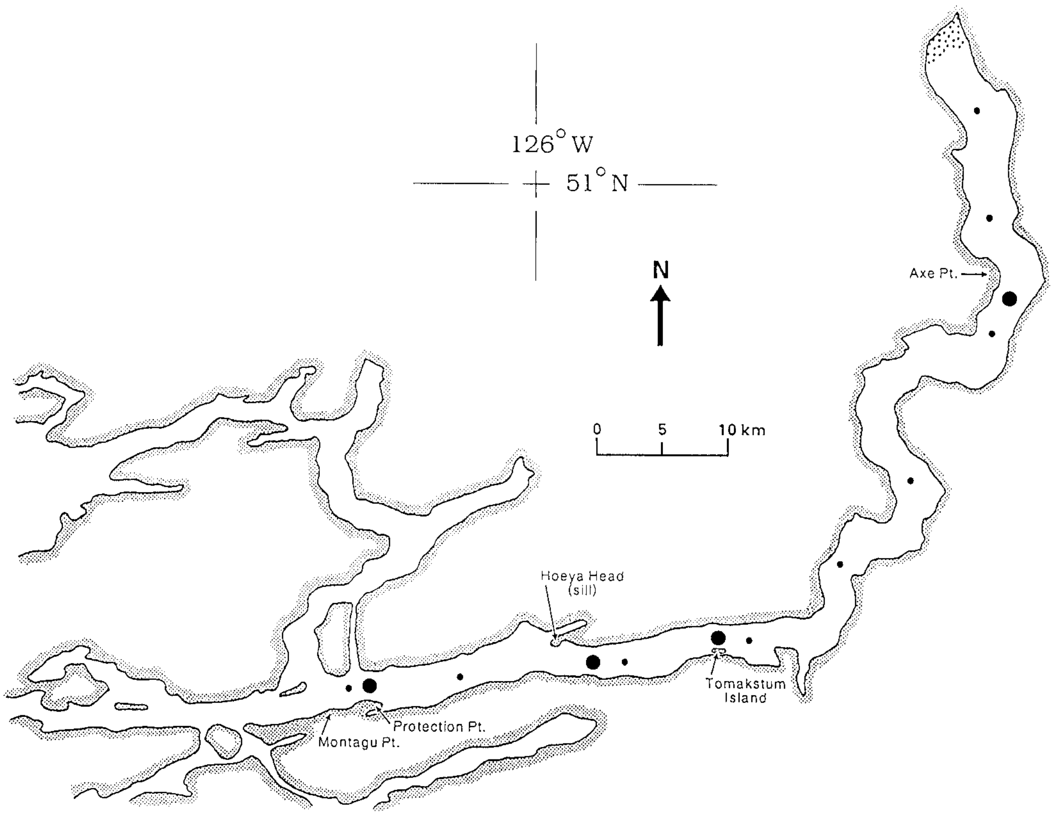

FIG. 1. Plan view of Knight Inlet [adapted from Stacey et al. (1995)]. The observations from the mooring near Protection Point (large solid circle) are located at the open boundary of the numerical model, and they are used to provide density input at the open boundary of the numerical model. The observations from the mooring near Tomakstum Island (large solid circle) are compared to the model simulations. The sill of the inlet is at Hoeya Head. The small solid circles show CTD stations.

ically realistic, particularly when wind forcing was present and surface waves should cause significant mixing near the surface. Consequently, Stacey and Pond (1997) modified the numerical model to take the wind mixing more explicitly into account, which simply meant modifying the 2.5 turbulence closure scheme of Mellor and Yamada (1982) by changing the surface boundary condition on the turbulent velocity scale q from that initially proposed by Mellor and Yamada (1982) to that of Craig and Banner (1994). Mellor and Yamada (1982) use the same boundary condition at the air–sea interface that they use at the air–solid interface:

$$q^2 = B_1^{2/3} u_*^2, \tag{1}$$

where $B_1 = 16.6$ and u_* is the friction velocity for the water. Craig and Banner (1994) use

$$\lambda_v \frac{\partial}{\partial z}\left[\frac{q^2}{2}\right] = \alpha u_*^3, \tag{2}$$

where λ_v is the vertical diffusion coefficient for the turbulent kinetic energy and α is a constant, set equal to 100 by Craig and Banner. By using (2), Stacey and Pond (1997) found that the simulation of the subtidal circulation in Knight Inlet was noticeably improved near the surface. It was also found that the simulated near-surface circulation was more sensitive to variations in z_0 (held constant for each simulation) when (2) was used. When (2) is used, z_0 has a direct influence on the surface flux of q^2 through its influence on λ_v.

Terray et al. (1996) present experimental evidence that α [see (2)] is not a constant but rather depends on the wave age. There is considerable scatter in the data presented by Terray et al. (1996), however, and for wave ages (C_p/u_{*a}, where C_p is the phase speed and u_{*a} the friction velocity for the atmosphere) greater than about 10, $\alpha \approx 150$ is consistent with the observations. This is the value for α that will be used in this note.

In this note, the model of Stacey et al. (1995) is used with boundary condition (2) to find what magnitude of z_0 gives the ‘‘best’’ simulation of the near-surface circulation. The sea state, on which z_0 presumably depends, was not directly observed, but it is known to depend on the wind speed that was observed (Fig. 2). Because the sea state was not directly observed and because observations were not made rapidly enough to resolve the velocity field of the surface waves, z_0 must be parameterized in terms of a length scale other than the

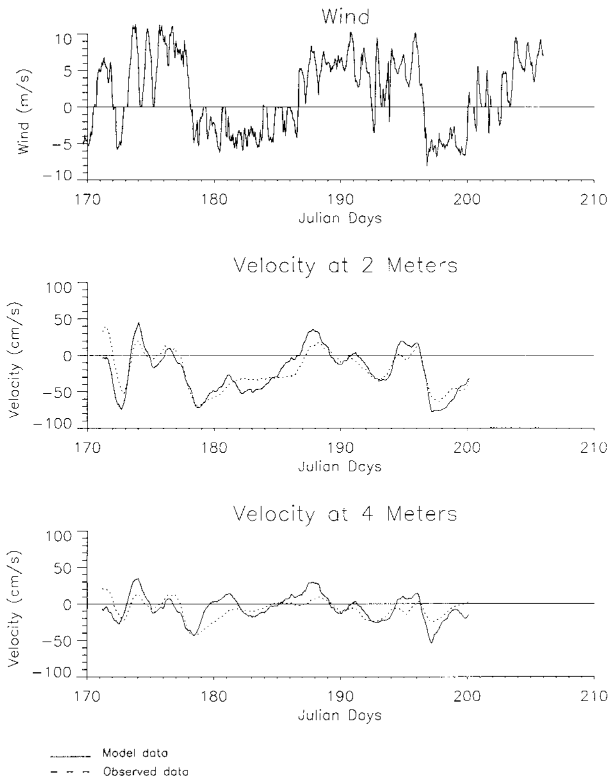

FIG. 2. (a) The observed along-channel wind speed 4 m above the water surface at Tomakstum Island. (b) The simulated (solid line) and observed (dashed line) subtidal along-channel velocity at 2 m. (c) Same as (b) but at 4 m. Positive values indicate motion in the up-inlet direction.

amplitude or wavenumber of the surface waves. This is a situation that modelers can expect to often find themselves in, so a formulation of z_0 in terms of more easily attainable information is desirable. A basic length scale that emerges for z_0 from scale analysis, and that has been successfully used on the atmospheric side of the air–sea interface (Charnock 1955), is

$$z_0 = a\frac{u_*^2}{g}, \tag{3}$$

where a is a constant and g is the acceleration due to gravity. The same expression has been proposed for z_0 on the water side of the air–sea interface, and Bye (1988) proposes that on the water side of the air–sea interface, a should take a value of about 1400 (when u_* is calculated using the density of water). When (3) is applied to the air side of the air–sea interface, a typically is given a value of about 10 (again using the density of water in the calculation of u_*). In order for (3) to have validity, surface waves should be breaking (i.e., the wind speed at 10 m should be about 5 m s^{-1} or greater), which is indeed the case for most of the duration of the experiment considered here. The winds in Fig. 2 are from 4 m, and the 10-m winds would be even stronger.

In this note, it will be shown that the most appropriate value for a, at least for the dataset examined here, is $a \approx O(10^5)$. When (3) is modified [see (5)] for comparison with a common empirical relationship for the significant wave height H_s [see (6) and (7)], one obtains $z_0 \approx O(H_s)$.

2. The model and the data

The model and the data have been described in detail elsewhere (e.g., Baker 1992; Baker and Pond 1995; Stacey et al. 1995; Stacey and Pond 1997), so a brief synopsis will suffice here. Moorings were deployed in Knight Inlet during 1988 and 1989, each time for a duration of about one month. The data used here will be those collected during 1989. The winds were less strong in 1988 (i.e., surface waves may not have been breaking during a significant portion of the experiment), and in 1988 the direction sensor in the anemometer at Tomakstum Island (Fig. 1) failed.

From mid-June until mid-July 1989, moorings were deployed at four locations in Knight Inlet (Fig. 1). In view of the work to be discussed here, the most important moorings were at Protection Point and Tomakstum Island, where observations using S4 current meters were made at depths 2, 4, 6, 9, and 12 m, and, using Cyclesondes and Aanderaa current meters, over the rest of the water column depth also. The mooring at Protection Point, which was located near the mouth of the inlet, provided input to force the numerical model at its open boundary, but the mooring at Tomakstum Island is completely independent of the model, so the data from there can be used to test the ability of the model to simulate the circulation in the inlet. Also, an anemometer was located near Tomakstum Island during the experiment, so the winds at the mooring are known. Another anemometer was located at Protection Point but it failed to work during the experiment. The along-channel velocity at 2 and 4 m will be used here to examine the dependence of the near-surface circulation on z_0. Because the tides as well as the winds are an important component of the circulation in the inlet and because we are interested in the wind-forced component of the near-surface circulation, all of the time series, simulated and observed, were filtered with a 25-h moving average in order to remove most of the tidal signal at diurnal periods and smaller before doing the basic statistical analyses. Once the tides have been removed, much of the remaining temporal variability in the near-surface velocity field will be caused by the winds.

The model is laterally integrated and uses the 2.5 turbulent closure scheme of Mellor and Yamada (1982). The diffusion coefficients for density, momentum, and q^2 are taken to be proportional to the product of q and the length scale l where near the surface we have the standard expression

$$l = k(z + z_0), \tag{4}$$

where $k = 0.4$ is von Kármán's constant, z is the distance from the surface, and z_0 is given by (3). The only other

TABLE 1. The sum of the square of the residuals (calculated after first removing the mean from each time series), the variance, and the mean for the subtidal velocity at Tomakstum Island. The observed and simulated values are given. Data from column 18 of the model are used to produce the simulated values given here. Stacey and Pond (1997) used column 17. One deduces the same ''best'' value for a when column 17 is used, but column 18 gives lower SSRs.

Depth (m)	a ($\times 10^5$)	SSR ($cm^2\ s^{-2}$)	Variance ($cm^2\ s^{-2}$)	Mean ($cm\ s^{-1}$)
2	Observed		632	−21.2
	0.1	236	1004	−20.8
	0.5	233	1010	−21.1
	1	238	1010	−21.4
	2	208	892	−21.9
	4	197	809	−23.0
	8	183	739	−23.6
	16	161	723	−25.6
4	Observed		165	−7.8
	0.1	218	284	−4.9
	0.5	218	285	−4.3
	1	219	303	−4.3
	2	187	297	−5.6
	4	194	328	−7.8
	8	198	371	−11.8
	16	244	411	−15.2

differences between the model used here and that used by Stacey et al. (1995) is that greater resolution is used here and boundary condition (2) is used instead of boundary condition (1). Here, there are 30 (10) grid points in the upper 10 (2) m.

3. The results

The model was run using a number of different values for a, and the basic statistical results are tabulated in Table 1. The values presented for a span the range of values that minimizes the sum of the square of the residuals between the observed and simulated data and that provide reasonable estimates of the mean. One way to estimate the ''best'' fit of the model to the observations is to scale the sum of the square of the residuals (SSRs) at 2 and 4 m by the observed variances and then to sum the scaled residuals to produce a single number. The best fit according to this criterion is attained when $a \approx 2 \times 10^5$, but the number does not vary very much for values of a between 2×10^5 and 8×10^5. Note also, however, that the simulated variance at 4 m begins to noticeably increase and diverge even further from the observed variance as a is increased to values larger than 2×10^5. Also, when $a \geq 8 \times 10^5$ the mean velocity at 4 m is not well simulated. When $a \approx 2 \times 10^5$–4×10^5 the mean velocity at 4 m is quite well simulated. Based on this extra information, one can deduce that $a \approx 8 \times 10^5$ is too large. When the SSRs and variances at 2 m are considered in isolation it appears that very large values of a, even greater than 8×10^5, give the best fit to the data, but in percentage terms the improvement at 2 m is less than the degradation at 4 m (and deeper depths not shown here). Also, the mean velocity at 2 m is not well simulated when $a \geq 8 \times 10^5$. It is clear that $a \approx O(10^6)$ is too large. It is harder to make a similarly definitive statement about the lower bound on a but it is likely that $a \approx O(10^4)$ is too small. Overall, $a \approx 2 \times 10^5$ gives about the best simulation although obviously this result is only good to within an order of magnitude.

In addition to $\alpha = 150$, other values for α were tried. It was found that for α between about 100 and 250 [its approximate upper bound, using the results of Terray et al. (1996)], a is still $O(10^5)$ although a (and therefore also z_0) must be decreased as α is increased. When $\alpha = 50$, the SSRs are larger overall, suggesting that α should be larger than 50 for the data from Knight Inlet. Craig (1996), for the situation he was considering, found the same qualitative relationship between z_0 and α, and calculated the detailed relationship between the two.

Figure 2 shows the time series of the simulated (when $a \approx 2 \times 10^5$) and observed subtidal velocity at 2 and 4 m. One sees that at 2 m in particular the model is providing an accurate simulation of the velocity. One also can easily see that the wind is having a very significant influence on the velocity at 2 m. (The down-inlet velocities tend to be larger than the up-inlet velocities because the estuarine circulation caused by freshwater runoff is in the down-inlet direction.) At 4 m the simulation is not as accurate, but the velocity is not as strong there, so perhaps this is not surprising. At around day 188, the simulated velocity is too large in magnitude at both 2 and 4 m, but the percentage discrepancy is larger at 4 m. At around day 180 the simulated velocity at 4 m undergoes a fluctuation that is absent in the observations.

4. Discussion and conclusions

For Knight Inlet in 1989, $a \approx O(10^5)$ is the most appropriate value, which is two orders of magnitude larger than the value of 1400 suggested by Bye (1988). Converting (3) into an expression that uses the wind speed (and using the drag coefficient $C_D = 0.0011$), one obtains

$$z_0 \approx 0.2\frac{W^2}{g} \quad (5)$$

when a equals 2×10^5. An empirical expression for the significant wave height H_s (e.g., LeBlond and Mysak 1980, p. 485) is

$$H_s \approx 0.3\frac{W^2}{g} \quad (6)$$

for unlimited fetch. Therefore one sees that according to these simulations and the empirical expression for H_s, $z_0 \approx O(H_s)$. The time-averaged value of z_0, for any simulation presented here using $a = 2 \times 10^5$, is $\overline{z_0} = 0.6$ m. The rms wind speed is 5.4 m s^{-1}. Pond and Pickard (1983) state that for a wind speed of 5 m s^{-1}, the wave field is almost fully developed and $H_s \approx 0.4$ m if the wind has been blowing for about 2 hours and the fetch is about 20 km. Tomakstum Island is about 60 km from the head of the inlet and the wind (Fig. 2) was often blowing for intervals of 2 hours or more.

Terray et al. (1996) and Drennan et al. (1996) do not make an explicit estimate of z_0, but they do consider H_s to be a significant depth scale below the surface layer. Also, when comparing their results to those of Craig and Banner (1994) they make the assumption that $z_0 \approx H_s$ in order to show that their results are consistent with the model of Craig and Banner. The results presented here support their supposition. The suggestion of Craig and Banner (1994) that z_0 is proportional to a wave amplitude is also supported by the results presented here.

For a wind speed of 4.7 m s^{-1} (close to the 5.4 m s^{-1} mentioned above), Craig (1996) finds that $z_0 = 0.01$ m, which is much smaller than the value of 0.45 m given by (5). However, as Craig notes, the waves observed in the laboratory are far from being fully developed. They are certainly farther from full development than the waves in Knight Inlet, so one would expect the roughness lengths caused by them to be much less than those caused by the waves in the fjord. It is certainly possible that the constant a in (3) should have a dependence on wave age, and that there exists a more appropriate length scale for z_0 than u_*^2/g, even though it is found to be an appropriate length scale on the air side of the air–sea interface. Certainly, (5) is inappropriate for very young waves as the results of Craig (1996) show, but in many cases one is interested in investigating wind-forced flows where the waves are more fully developed than in the case examined by Craig. When one applies (5) to the SWADE data from the North Atlantic Ocean (Drennan et al. 1996) where the wind speed at 12 m was about 10 m s^{-1}, one obtains a roughness length of about 2 m. Drennan et al., using Craig and Banner's model, estimate the roughness length to be 1–3 m.

In a recent paper, Gemmrich and Farmer (1998) suggest that $z_0 \approx 0.2$ m, independent of H_s, which during their experiment had a mean value of 4.5 m. They note that their estimate for z_0 is comparible in magnitude to the vertical scale of air entrainment by whitecaps. Gemmrich and Farmer's estimate for z_0 is of the same order of magnitude as the mean estimate given here, so it may be that the estimate presented here supports their result and that it is just a coincidence that the estimated z_0 for Knight Inlet is about the same as H_s. It is also possible that z_0 approximates H_s at the moderate wind speeds observed in Knight Inlet but not at the larger wind speeds (13–17 m s^{-1}) observed by Gemmrich and Farmer. The longer wavelength waves that exist at the larger wind speeds may not influence the value of z_0. Or, as Gemmrich and Farmer (and others) mention, the value of the dimensionless coefficient S_q used in the expression for the diffusion coefficient λ_v for q^2 (i.e., $\lambda_v = S_q lq$), which Mellor and Yamada (1982) set equal to 0.2, may require adjustment when the diffusion of interest is occurring in the presence of surface waves. Any adjustment of S_q might require an adjustment of z_0 also.

Finally, even though uncertainties remain about how best to express z_0 (and other parameters), as the time series of the near-surface, subtidal velocity (Fig. 2) show, a numerical model that uses the Mellor–Yamada turbulence closure scheme can simulate the near-surface, wind-forced velocity field quite well.

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