A Time-Dependent Model of a Coastal Polynya*

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ABSTRACT

An idealized model is used here to examine the temporal behavior of a coastal polynya driven by an offshore wind. The model has incorporated the important effect of a finite surface drift by which the frazil ice formed in the open ocean is advected downwind and collected at the ice edge. It is found that a finite surface drift, on the one hand, reduces the "inertia" of the ice edge, thus prompting a greater response for a given flux inbalance, but, on the other, delays and smooths out the forcing effect at the ice edge. The resulting polynya behavior is examined in response both to a finite step change and small perturbations in the atmospheric forcing. The model results suggest, among other things, that the ice edge is in approximate equilibrium with synoptic and longer-period atmospheric variations, but is not responsive to higher-frequency atmospheric fluctuations with periods short compared with the transit time of a water parcel through the polynya.

1. Introduction

In polar seas, even in the dead of winter, open areas (polynyas) are frequently observed between the coast (or land fast ice) and the ice pack when there are prevailing offshore winds (Knapp, 1972). Because of the shallowness of the coastal ocean, the water is well mixed and near the freezing temperature. As a consequence, the oceanic heat flux can at most supply a fraction of the heat loss to the atmosphere which, without insulation of the ice cover, can reach values as high as 500 W m^{-2} (Cavalieri and Martin, 1985). It is thus generally believed that ice is rapidly produced in these areas and it is the mechanical removal of the ice by the offshore wind that keeps the areas open. This mechanism is clearly suggested in the photograph taken south of Nome, Alaska (Fig. 1, reproduced from Martin and Kauffman, 1981) that shows streaks of grease ice (a slurry of frazil ice platelets) herded downwind in Langmuir cells and piled up against ice floes.

Since the total ice produced in a polynya is proportional to its area, Lebedev (1968) argued that the ice edge cannot move downwind indefinitely but must reach an equilibrium when the offshore ice flux just balances the ice production. He further deduced that, because of the countering effect of a varying wind speed on the ice drift and ice production, particularly during moderate to high wind regimes, the polynya width does not depend strongly on the wind speed but is primarily a function of the air temperature. Pease (1987) has additionally examined the onset of a polynya under a steady wind and found that the adjustment time is scaled by the ratio of the ice thickness at the ice edge to the ice production rate over the open ocean, which she estimated to be of the order of a few days or less.

One simplifying assumption made by Pease is that the surface drift over the open ocean is infinite so that the newly formed ice is instantaneously deposited at the ice edge. This is at odds with observations that surface drift speed is comparable to that of the ice drift. Indeed, as a percentage of the wind speed measured 10 m above the surface, values of 3.5% for the surface drift and 2% for the ice drift have been frequently cited in the literature as general rules (e.g., see Amstutz and Samuels, 1984; Zubov, 1943; McPhee, 1980). It is thus important to examine the effect of a finite surface drift on the temporal behavior of the polynya.

To this aim, we consider a highly idealized model as formulated in section 2. The polynya response to a finite step change and small perturbations in the atmospheric forcings are discussed respectively in sections 3 and 4. Some model assumptions are discussed in section 5 and the main results of the model are summarized in section 6.

2. The model

Let us consider a coastal polynya driven by an offshore wind (u_A) as schematically shown in Fig. 2. The heat loss to the atmosphere from the open ocean causes production of the frazil ice (at a rate P), which is "herded" downwind at a surface drift speed u_W. In the

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FIG. 1. Oblique aerial photograph from 150 m of a polynya south of Nome, Alaska on 5 March 1978 (reproduced from Martin and Kauffman, 1985). The floe diameter in the center of the photograph is about 100 m. The air temperature was −20°C and the wind speed was 15 m s^{-1} directed toward the floes.

mean time, the consolidated ice is advected offshore at the ice drift speed u_I. For simplicity, we shall consider a polynya driven by synoptic winds (with scales of hundreds of kilometers) so that the atmospheric variables are spatially uniform over the polynya. Consequently, the ice production rate P and the surface drift speed u_W are assumed spatially uniform and functions of time only. The effect of a more localized katabatic wind will be discussed in section 5. For simplicity, we shall further assume that u_W and u_I are linearly proportional to the instantaneous wind (see section 5) so that the ratio $k \equiv u_I/u_W$ is a positive constant. This constant is further restricted to be smaller than one, based on the discussion in the previous section.

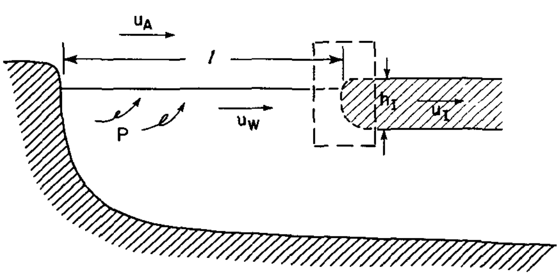

FIG. 2. A schematic of a coastal polynya driven by an offshore wind (u_A). The frazil ice is produced in the open ocean at a rate P and herded downwind at the surface drift speed u_W while the consolidated ice is advected offshore at the ice drift speed u_I. The dashed box indicates the control volume encompassing the ice edge which has a coordinate l (measured from the coast). h_I is the thickness of the consolidated ice exiting the control volume.

To derive the flux balance at the ice edge, let us consider a control volume (the dashed box in Fig. 2) encompassing the ice edge where active collection and solidification of frazil ice occur. Since the ice edge is rather well defined based on Fig. 1, we can choose the control volume to be sufficiently narrow that its position is specified by the coordinate l (measured from the coast or land fast ice) and that the temporal variation of ice mass within the control volume is small compared with ice fluxes across the box. The ice flux into and out of the control column must then balance.

Let h be the amount of ice contained in a water parcel (of unit surface area) arriving at the ice edge, then

$$h = \int_{t-\Delta t}^{t} P(t)dt, \tag{1}$$

where Δt is the transit time of the water parcel through the polynya and P is the ice production rate in units of ice thickness per unit time per unit surface area. The transit time is in turn related to the polynya width l by the expression

$$l = \int_{t-\Delta t}^{t} u_W(t)dt. \tag{2}$$

The flux balance, viewed from the moving frame affixed to the ice edge, states that

$$h_I\left(u_I - \frac{dl}{dt}\right) = h\left(u_W - \frac{dl}{dt}\right), \tag{3}$$

where h_I is the thickness of the consolidated ice exiting from the control volume. For simplicity and for lack of understanding of the frazil collection process at the ice edge, h_I is taken to be constant (see further discussions in section 5). Given atmospheric forcing functions (and hence P and u), Δt can in principle be eliminated from (1) and (2) to yield a functional $h(l, t)$, and (3) can be solved for $l(t)$.

For the special case of a steady forcing, (1) and (2) imply

$$h = P\Delta t = Pl/u_W, \tag{4}$$

which, when substituted into (3), yields

$$h_I\left(u_I - \frac{dl}{dt}\right) = \frac{Pl}{u_W}\left(u_W - \frac{dl}{dt}\right). \tag{5}$$

Setting $dl/dt = 0$, the steady state solution is given by

$$l = l_S \equiv h_I u_I/P. \tag{6}$$

This is the maximum lead width derived by Lebedev (1968) and discussed in some detail by Pease (1987), which simply states the balance between the offshore ice flux driven by wind and the total ice production over the polynya.

Assuming an infinite surface drift speed (i.e., $u_W \gg dl/dt$), Pease has solved (5) for the onset of a polynya under a steady forcing. As can be seen from the equation, the ice edge moves offshore with the ice drift u_I initially and decelerates exponentially toward the equilibrium with an e-folding time of h_I/P. But in view of the observations noted above that the surface drift is comparable to the ice drift, it is not consistent to neglect the ice edge motion with respect to the surface drift while retaining it with respect to the ice drift. In fact, as we shall see later, during the shoreward retreat of the ice edge, the ice edge speed has no upper bound. It is therefore necessary to include the last term in (5) for a more realistic modeling of the temporal behavior of the polynya.

3. Finite adjustment

We shall first examine the polynya response to a step change of forcing at, say, $t = 0$. The forcing before and after the change will be referred to as "old" and "new", respectively. To simplify the mathematics, the variables are nondimensionalized by the scaling rules (indicated by brackets) based on the new forcing: $[h] = h_I$, $[u] = u_I$, $[l] = l_S$, and $[t] = l_S/u_I \equiv h_I/P$. Equation 3 then becomes,

$$1 - \frac{dl}{dt} = h\left(\frac{1}{k} - \frac{dl}{dt}\right), \tag{7}$$

where $k \equiv u_I/u_W$ is a positive constant smaller than one. It is seen that k is the only internal parameter that governs the temporal behavior of the polynya.

Let l_0 be the initial location of the ice edge in equilibrium with the old forcing, we shall seek to determine the evolution of l from l_0 to its new equilibrium value of one as schematically shown in Fig. 3. Let ζ be the distance over which the water parcel arriving at the ice edge at time t has traveled under the old forcing, then, noting that the surface drift has a dimensionless speed $1/k$,

$$\zeta = l - t/k. \tag{8}$$

The distance ζ decreases from its initial value of l_0 to a value of 0 at $t = t_C$, defined as the time after which all the subsequently arriving water parcels have been exposed only to the new forcing.

For $t > t_C$, h is then given by [see Eq. (4)]

$$h = kl, \tag{9}$$

and (7) becomes

$$1 - \frac{dl}{dt} = kl\left(\frac{1}{k} - \frac{dl}{dt}\right). \tag{10}$$

This equation can be linearized by the transformation of variables: $l(t) \rightarrow t(l)$, which yields

$$\frac{dt}{dl} = \frac{1 - kl}{1 - l}. \tag{11}$$

The first integral gives

$$t = t_C + k(l - l_C) - (1 - k)\ln\frac{1 - l}{1 - l_C}, \tag{12}$$

where l_C is the coordinate of the ice edge at $t = t_C$. The

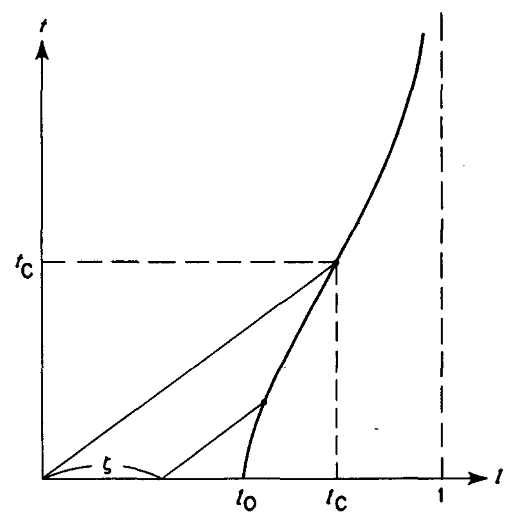

FIG. 3. A schematic showing polynya width (l) as a function of time when the forcing undergoes a step change at $t = 0$. l_0 is the initial position of the ice edge in equilibrium with the old forcing and ζ is the distance over which the water parcel arriving at the ice edge at time t has been exposed to the old forcing. (l_C, t_C) indicates the transition between the "runaway" and the "approach" phase of the adjustment.

constants l_C and t_C need to be determined from the solution for $t < t_C$.

For $0 < t < t_C$, h consists of ice produced both under the new and old forcings. The ice produced under the new forcing is proportional to time t or, in dimensionless unit, given exactly by t. The ice produced under the old forcing is proportional to ζ, the distance and hence the time over which a water parcel has travelled under the old forcing. Since this component has the equilibrium value of k [see Eq. (4)] at $t = 0$ (when $\zeta = l_0$), it is given by $k\zeta/l_0$. The total amount of ice arriving at the ice edge is thus

$$h = t + k\zeta/l_0 \quad \text{for} \quad 0 < t < t_C, \tag{13}$$

which, when substituted into (7) and using (8), leads to

$$\frac{d\zeta}{dt} + \frac{1}{\gamma} = \left(t + \frac{k\zeta}{l_0}\right)\frac{d\zeta}{dt}, \tag{14}$$

where $\gamma \equiv k/(1-k)$. The equation can again be linearized by the transformation $\zeta(t) \rightarrow t(\zeta)$, which gives

$$\frac{dt}{d\zeta} - \gamma t = \gamma\left(\frac{k\zeta}{l_0} - 1\right). \tag{15}$$

The solution subject to the initial condition $t(l_0) = 0$ is

$$t = k\left\{1 - \frac{\zeta}{l_0} + \frac{1}{\gamma}\left(1 - \frac{1}{l_0}\right)\left[1 - e^{-\gamma(l_0-\zeta)}\right]\right\}. \tag{16}$$

Combined with (8), the solution $l(t)$ can be calculated for $t < t_C$. Setting $\zeta = 0$ in (16) and (8), we derive that

$$t_C = k\left\{1 + \frac{1}{\gamma}\left(1 - \frac{1}{l_0}\right)\left[1 - e^{-\gamma l_0}\right]\right\}, \tag{17}$$

$$l_C = \frac{1}{k}t_C, \tag{18}$$

which then specify the solution (12) for $t > t_C$.

To illustrate the polynya behavior, solutions for $l_0 = 0$, ½ and 2 for selected values of k are plotted in Fig. 4. For the special case of $l_0 = 0$ or $k = 0$ (i.e. infinite surface drift), (17) implies $t_C = 0$ and the solution is given by (12) with $l_C = 0$ and l_0, respectively. It is seen that during the onset of the polynya (i.e. $l_0 = 0$), the finite surface drift ($k > 0$) acts to speed up the forward motion of the ice edge by reducing the efficiency of the ice collection at the ice edge.

For the finite adjustment between two polynya states ($l_0 \neq 0$), however, the polynya behavior is qualitatively modified by the finite surface drift. When the surface drift is infinite, there is an instantaneous collection of all the newly formed ice at the ice edge which thus responds instantly to the forcing change and then decelerates toward the new equilibrium. But when the surface drift is finite, the amount of ice arriving at the ice edge, being accumulated through the transit of the open ocean, varies continuously in time. The ice edge

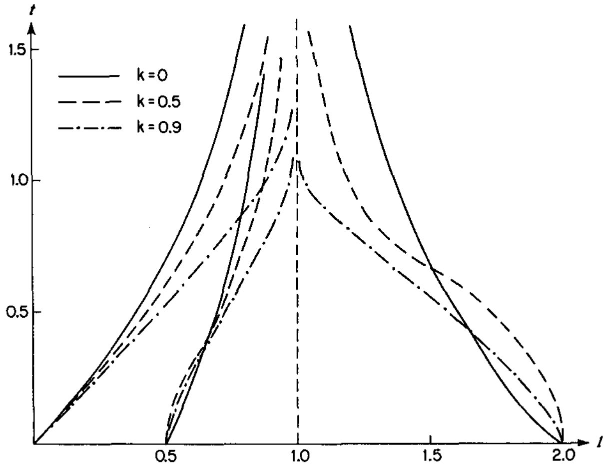

FIG. 4. Polynya width (l) as a function of time in response to a step change of forcing at $t = 0$ for the case of $l_0 = 0$, ½, 2 and $k = 0$, 0.5 and 0.9.

thus remains stationary initially and accelerates in speed as the newly arrived water parcels bring in an amount of ice (h) that increasingly deviates from its initial value (since they have been exposed to new forcing over longer periods of time). The ice edge attains a maximum speed at $t = t_C$ after which it decelerates in speed as h approaches the new equilibrium. The adjustment thus consists of two phases: the initial "runaway" phase when the ice edge is set into motion by disturbances of the initial equilibrium which accelerates in speed and the subsequent "approach" phase when the ice edge decelerates toward the new equilibrium.

Despite this two-phase adjustment, the final equilibrium is reached sooner when the surface drift is finite and increasingly so as the surface drift speed decreases (with respect to that of the ice drift). In fact, one can show that t_C and l_C [Eqs. (17) and (18)] approach one as k approaches one, regardless of the value of l_0. That is, when the surface drift speed approaches that of the ice drift, the ice edge is near its final equilibrium within one time unit, regardless of the distance of the adjustment.

One also notices that although during the forward adjustment the ice edge speed is bounded by the ice drift, there is no such limit during the backward adjustment when the ice edge retreats to the coast. Indeed, in this latter case, as plotted in Fig. 5, the maximum ice edge speed (i.e. at $t = t_C$) can reach great values when the distance of adjustment is large. To help understand these results, we rewrite the governing equation (7) as

$$(1 - h)\frac{dl}{dt} = 1 - \frac{h}{k}. \tag{19}$$

The right hand side (rhs) is the flux deficit at the ice edge that prompts its motion, the magnitude of which however is regulated by the "inertia" $1 - h$, the differ-

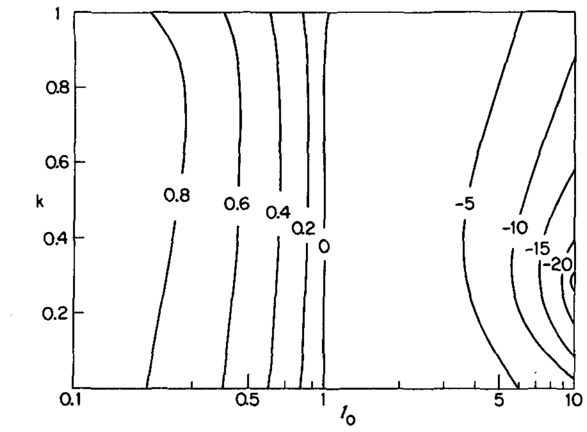

FIG. 5. The maximum ice edge speed (at $t = t_C$) as a function of l_0 and k.

ence of ice thickness across the ice edge. Obviously the smaller this difference, the greater the ice edge motion is needed to counter any given flux inbalance. For an infinite surface drift, h vanishes [see Eq. (9)] and this inertia is unity, but for a finite surface drift, this inertia is variable and smaller than one. Since during the approach phase ($t > t_C$), the rhs is $1 - l$ [see Eq. (9)], the ice edge associated with a finite surface drift has greater speed for a given departure from the final equilibrium, as is clearly seen in Fig. 4. The contrast between the forward and backward adjustment between two polynya states is also evident from (19). For the forward adjustment, h decreases from its initial value k to some minimum value at the end of the runaway phase before it increases back to the final equilibrium value k, dl/dt thus is positive and bounded by unity. For the backward adjustment, however, h varies in the opposite sense, i.e. it increases from k to some maximum value (smaller than one because of the positive feedback between l and h), and then decreases back to k; dl/dt thus has no particular upper bound and can reach great values. It is also interesting to note from Fig. 5 that for a given l_0, the maximum ice edge speed during the backward adjustment has a maximum at some intermediate value k, a result of intricate balance between the ice flux surplus and the ice edge inertia.

In conclusion, a finite surface drift accelerates the adjustment of the ice edge to its final equilibrium, which is generally accomplished within the time scale h_I/P or, according to estimates of Pease (1987), of the order of a few days or less. Thus for synoptic and longer period atmospheric variations, the polynya is in approximate equilibrium with the atmospheric forcing and its width can be reasonably described by the steady state solution of section 2.

4. Small perturbations

In this section, we shall examine the polynya response to small amplitude atmospheric variations. Using scaling rules of section 3, but based on mean conditions, the nondimensionalized variables are given by

$$(P, u, l, \Delta t/k, h/k) = 1 + \epsilon(P', u', l', \Delta t', h'), \quad (20)$$

where the primed variables are time-dependent perturbations and ϵ is assumed much smaller than one. Substituting (20) into the nondimensionalized version of (1), (2) and (3), the leading order balances are given by

$$h' = \Delta t' + \frac{1}{k}\int_{t-k}^{t} P'dt', \quad (21)$$

$$\Delta t' = l' - \frac{1}{k}\int_{t-k}^{t} u'dt', \quad (22)$$

$$(1 - k)\frac{dl'}{dt} = -h'. \quad (23)$$

Of particular significance, we observe that u' does not enter (23) directly. This is because the wind speed variation affects similarly the surface and ice drift and thus does not directly offset the flux balance at the ice edge. The adjustment is thus prompted only by variation of h, the amount of ice arriving at the ice edge. Equation (21) states that h varies with the transit time of a water parcel through the polynya and the rate by which the ice is being produced and accumulated during this transit. Equation (22) states that the transit time, in turn, varies with the distance and speed by which a water parcel has traveled.

Substituting (21) and (22) into (23), we derive that

$$(1 - k)\frac{dl'}{dt} + l' = \frac{1}{k}\int_{t-k}^{t} (u' - P')dt' \equiv F(t), \quad (24)$$

which has the solution

$$l'(t) = e^{-(1-k)t}\int_0^t e^{(1-k)t'}F(t')dt' + l'(0). \quad (25)$$

Given atmospheric variations (u' and P'), l' can then be calculated subject to any initial condition $l'(0)$.

Some general properties of the polynya response can be inferred from (24) without any specific calculations. We first notice that there are two intrinsic time scales in the equation: the relaxation time $(1 - k)$ associated with the finite inertia of the ice edge, and the transit time (k) by which a water parcel travels through the polynya. Of particular significance, the atmospheric perturbations enter only through its averaged effect over this transit time. Thus, for atmospheric variations with periods equal to an integral fraction of, or short compared with, the transit time, the net forcing (F) experienced at the ice edge is likely to be small. In addition to this smoothing effect, the finite inertia also tends to reduce the polynya response to high frequency atmospheric variations: It can be seen from (24) that for forcing periods short compared with the relaxation time, the response amplitude is proportional to the forcing period. One thus concludes that, because of the smoothing effect on the atmospheric perturbations and the finite inertia of the ice edge, the ice edge is not responsive to high frequency atmospheric variations, especially when the periods are short compared with the transit time of a water parcel through the polynya.

Some effects of varying k can also be inferred from (24). Increasing k (i.e., slowing down the surface drift with respect to the ice drift), on the one hand, lengthens the transit time and further smooths out the forcing perturbations, but, on the other, reduces the inertia of the system and prompts a faster and greater response of the ice edge.

For low frequency atmospheric variations with periods long compared with both the relaxation and the transit time, (24) reduces to $l' \approx u' - P'$, which can also be derived by perturbing the steady state solution (6). It is interesting to notice that the forcing enters only through the difference of u' and P', the former being through the wind effect on the surface drift and hence the transit time of a water parcel in the polynya [see Eq. (22)]. Since a varying wind speed also affects P' proportionally, particularly during moderate to high wind regimes, its net effect on l' is small; i.e., perturbations in wind speed are not effective in eliciting polynya response. On the other hand, perturbations in the air temperature affect only the ice production rate and hence are more effective in generating ice edge motions. The above results are consistent with the observations along the Antarctic Wilkes Land Coast (Cavalieri and Martin, 1985) where only marginal correlation between the polynya area and wind speed was detected. In addition, the largest polynya event there (around their Julian day 222) did not concur with appreciable change of wind speed, but followed a considerable warming of the atmosphere.

5. Discussions

We have extended Pease's model (1987) to include the effect of a finite surface drift in addressing the polynya response to atmospheric forcing. To isolate this effect, and for mathematical simplicity, we have however neglected other physical elements that are also important in simulating the polynya behavior. Some of the simplifying assumptions are discussed below.

We have assumed that the ice drift is linearly proportional to the instantaneous wind while more realistically the ice motion lags the wind by some appreciable interval. Simple scaling arguments suggest a lag of $t_L \approx \rho_I h_I/(\rho_W C_W U)$ [e.g., see Eq. (5) of Pease et al., 1983] where ρ_I and ρ_W are the ice and water density, respectively, h_I is the ice thickness, C_W is the water drag coefficient, and U is the relative ice-water speed. Using values of $\rho_I \approx 0.91 \times 10^3$ kg m^{-3}, $\rho_W \approx 10^3$ kg m^{-3}, $h_I \approx 0.1$ m (Pease, 1987), $C_W \approx 5.5 \times 10^{-3}$ (e.g. McPhee, 1980), and $U \approx 0.1$ m s^{-1}, t_L is less than an hour. The model results obviously need to be modified for time scales shorter than t_L. For example, we have argued that wind speed affects similarly the ice and surface drift and hence does not alter the flux balance that prompts the ice edge motion. But because of this lagged response of ice relative to the surface drift, a wind increase is expected to cause some shoreward motion of the ice edge initially before the ice edge returns to the equilibrium position.

Another simplifying assumption is that the thickness of the consolidated ice (h_I) is constant. Parameterization of h_I requires some understanding of the frazil collection process at the ice edge. Although there have been laboratory (Martin and Kauffman, 1981) and model (Bauer and Martin, 1983) studies of the grease ice growth in small leads, their results are not directly applicable to polynya where the collection process, being dominated by Langmuir circulation, is quite different. The lack of understanding of this process does not warrant its incorporation in the model.

We have assumed, for simplicity, that the atmospheric variables are spatially uniform over the polynya. This is not the case for polynyas driven by katabatic winds that weaken rapidly offshore (Kurtz and Bromwich, 1985). Some qualitative effects of a steady katabatic wind however can be assessed. We first notice that, since the wind affects similarly the ice production rate and the surface drift, the ice accretion per unit distance downwind (P/u_W) generally varies more gradually than the wind itself. In fact, in the case of moderate to high winds when the heat loss and hence ice production rate over the open ocean is linearly proportional to the wind speed, P/u_W is primarily a function of air temperature, which generally varies on the larger synoptic scales. In this case,

$$h = \int_0^l P/u_W dx \approx Pl/u_W,$$

the same expression as (4), and the polynya width is as given by (6), which is independent of the spatial variation of the wind speed. In practice, due to heat loss associated with free convection, radiation, etc., the ice production rate remains finite even when wind vanishes. The ice accretion rate per unit distance thus increases downwind and becomes infinite where the wind vanishes. The polynya width must then be bounded by the offshore extent of the katabatic wind and can be considerably smaller if the relative heat transfer, not directly driven by the wind shear, increases.

6. Summary

Through an idealized model, we have examined the temporal behavior of a coastal polynya driven by an offshore wind, particularly its response to a finite step change and small perturbations in atmospheric forcings. The model has incorporated the previously neglected effect of finite surface drift, which is found to have important ramifications on the polynya behavior.

During the onset of a polynya, a finite surface drift only acts to speed up the downwind motion of the ice edge, which decelerates monotonically toward the equilibrium. But during adjustment between two polynya states, a finite surface drift qualitatively modifies the polynya behavior and results in a two-phase adjustment: an initial "runaway" phase when the ice edge is set into motion by perturbation of the initial equilibrium that accelerates in time, and a subsequent "approach" phase when the ice edge decelerates toward the final equilibrium. The maximum ice edge speed at the transition of the two phases is bounded by ice drift during the forward adjustment, but can reach large values during the backward adjustment when the distance of adjustment is large.

As the surface drift speed decreases with respect to the ice drift, the adjustment to the final equilibrium is more complete within the adjustment time scale of h_I/P where h_I is the ice thickness at the ice edge and P is the ice production rate over the polynya. Using typical values, this time scale has been estimated by Pease (1987) to be of the order of a few days or less. A polynya is thus likely to be in approximate equilibrium with synoptic or longer-period atmospheric variations and its width reasonably described by the steady state solution.

For small-amplitude atmospheric perturbations, it is found that the ice edge responds only to the averaged effect of atmospheric perturbations on the arriving water parcels. Coupled with the finite inertia of the ice edge, it is thus not responsive to high-frequency atmospheric fluctuations with periods short compared with the transit time of the water parcel through the polynya. Because of the countering effect of wind speed on the transit time and the ice production rate over the polynya, particularly during moderate to high wind regimes when the ice production rate varies linearly with the wind speed, perturbation in wind speed is not effective in eliciting polynya response. Perturbation in the air temperature, on the other hand, affects only the ice production rate, and is thus more effective in generating ice edge motions.

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