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AUTOMATIC MEASUREMENTS OF VERTICAL OCEAN HEAT FLUX AND
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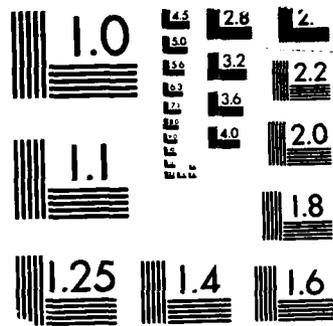
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**Automatic Measurements of
Vertical Ocean Heat Flux and Ice Mass Balance**

by

N. Untersteiner and A. S. Thorndike

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SUMMARY

The exchange of heat between ocean and floating sea ice is an unknown component of the heat balance of polar oceans. No direct measurements by either eddy flux or profile methods of useful duration have been made, and only a few sporadic estimates exist which derive ocean heat flux as a residual in the heat and mass balance of the underside of the ice. This lack of data is a serious impediment to developing dynamic-thermodynamic sea ice models (Parkinson and Washington, 1979; Hibler, 1980; Pollard et al., 1983) which would make it possible to adjust the ocean heat flux to achieve a realistic simulation of the mean annual cycle of sea ice extent.

We describe a new type of data buoy capable of making automatic and unattended measurements of the vertical flux of sensible heat in the oceanic boundary layer under a slab of floating ice. The method would also yield data on snow fall, surface melting and runoff, and bottom melting or freezing, i.e., the mass balance of the floating ice. The hardware consists of well-tested and proven components: pressure sensors, thermistors, solid-state switches, and the Argos positioning and data transmission link.

The new instrument should provide a cost effective way to acquire needed data that otherwise can only be obtained at manned drifting ice camps. The new data buoy should be suitable for use in both the polar pack ice and the rapidly melting ice of the marginal ice zones, provided that the ice is not thinner than approximately 1 m.

1. INTRODUCTION

Sea ice changes its thickness by melting or freezing at its top and bottom surfaces in response to the balances of heat fluxes at these surfaces. Measurements of the changes in ice thickness can therefore be interpreted as integrals of the surface heat balances. We propose a new, automatic measurement system enabling a program to monitor these ice thickness changes and heat balances. Development of this system would take about 2 years, with the first year devoted to building and testing the instruments and the second to field observations. The results would be expected to improve our understanding of the atmospheric and oceanic surface heat balances in the Arctic, and of the annual cycle of freezing and melting.

The strengths of this approach are seen in the simplicity of the measurement system and the promise of a long record of useful air-sea-ice data. Costs could be minimized by coordinating the program with other projects such as the Arctic Basin Buoy Program.

2. THE HEAT BALANCE AT THE LOWER SURFACE

When heat loss to the atmosphere has caused a solid layer of ice to form at the ocean surface, the further growth of that layer is controlled by the difference between heat conduction in the ice and turbulent transfer of sensible heat in the ocean:

$$\rho_i q \frac{dH}{dt} = k_i \frac{dT_i}{dz} - F_w . \quad (1)$$

where ρ_i , q , H , k_i , and T_i are the density, heat of fusion, thickness, thermal conductivity, and temperature of the sea ice. F_w is the flux of heat from the ocean to the bottom of the ice, and z is the vertical coordinate.

If the ice cover remains in a region of more or less uniform climate, it attains an equilibrium thickness when summer ablation is balanced by winter accretion. In the Central Arctic, this equilibrium thickness is between 3 and 4 m, a figure established by numerous observations from drifting stations as well as by upward-looking sonar profiles taken from submarines. The thermodynamic sea ice model developed by Maykut and Untersteiner (1971) reaches this equilibrium thickness when the vertical heat flux in the ocean is assumed to be between 1 and 2 W m^{-2} . That number is also obtained by applying Eq. (1) to direct observations of ice thickness and ice temperature (Untersteiner and Badgley, 1958; Untersteiner, 1961; Badgley, 1966), leaving the oceanic heat flux as a residual. The same approach was later adopted by Weller (1968) and Allison (1979) in studies of fast ice near the Antarctic coast.

The sensitivity of the ice thickness to the oceanic heat flux F_w can be judged from Fig. 1 which shows that, within a range of only several watts per square meter, F_w causes the equilibrium ice thickness to change by several meters. Despite several attempts (e.g., Lake, 1967), *long-term* observations of F_w by either the profile or the eddy flux methods have been unsuccessful and are probably infeasible.

What little information is available about F_w comes from a few time series of ice temperature and thickness taken at drifting stations. A recent example of such data is shown in Fig. 2. The method of using them is outlined in the next section.

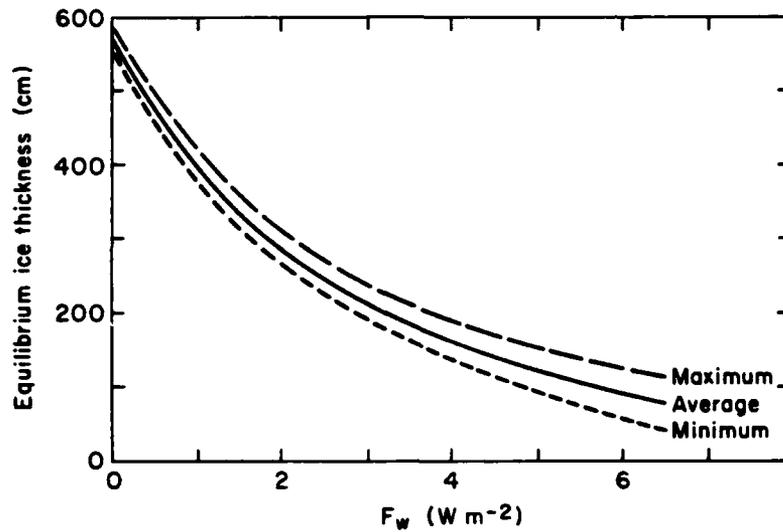


Figure 1. Equilibrium ice thickness as a function of the oceanic heat flux (from Maykut and Untersteiner, 1971).

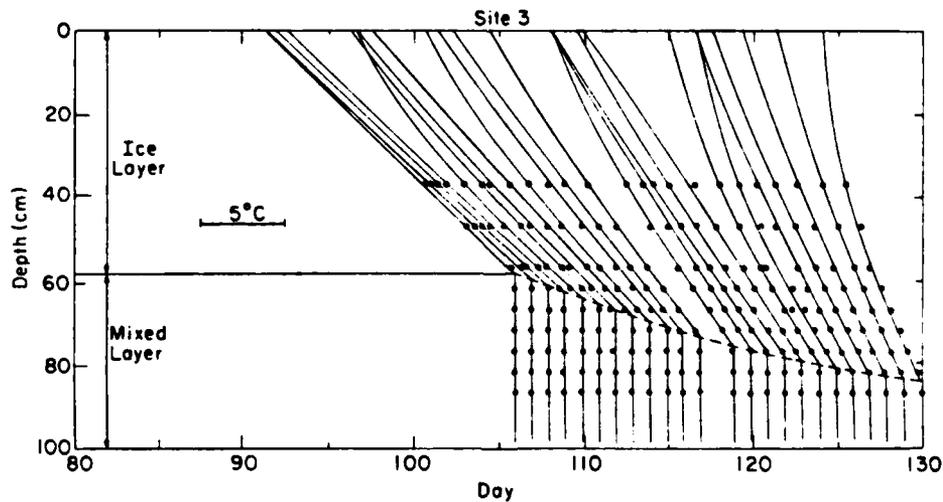


Figure 2. Temperature measurements across the underside of growing ice, obtained at drifting station FRAM 1 north of Fram Strait in spring of 1979 (from McPhee and Untersteiner, 1982). Profiles are fitted to data points daily. The reference temperature of the isothermal mixed layer beneath the ice was $\sim 1.75^\circ\text{C}$.

3. MEASURING CHANGES OF ICE THICKNESS

Unlike direct observations of heat flux in the ocean, measuring mass changes at the upper and lower surfaces of sea ice is easy, though laborious. It requires the reading of a number of ablation/accumulation stakes (enough to eliminate local scatter) on the top, and the frequent drilling of holes in the ice to follow ablation or accretion at the bottom.

An increment of automation was introduced by Untersteiner (1961, see Fig. 3), who used a water level recorder in combination with a frozen-in thickness gauge made of wire that, once emplaced, could be released by electrical heating. To measure thickness, the wire was pulled up until a cross bar on the end was against the bottom of the ice, and its thickness determined by comparing the height of the wire against a scaled stake frozen in the top of the ice. This eliminated the need for repeatedly drilling holes in the ice. Schwerdtfeger (1968) designed and built an automated version of that thickness gauge and used it successfully on fast Antarctic ice.

The new system described here is derived with modern techniques from the one discussed in the previous section. Its basic design is sketched in Fig. 4. The critical sensors are a string of thermistors to measure the profile of temperature across the bottom surface of the ice, and a differential pressure sensor. One side of this pressure sensor is in contact with the ocean; the other is in contact with the atmosphere through a vent running up the inside of the buoy tube. The entire assembly is fixed with respect to the ice by an anchor flange or similar device.

If a hole were drilled through the ice, seawater would rise in the hole to the freeboard level. The differential pressure sensor measures the depth, $H-h+d$, of the sensor below that level: $P = P_{ocean} - P_{air} = \rho_w \cdot g \cdot (H-h+d)$, where g is the acceleration due to gravity. Changes in pressure, P , correspond to changes in $H-h+d$; these changes occur as the ice slab grows or melts, as follows:

- (1) A layer of ice thickness δH , formed at the bottom surface, causes the slab to rise a height $\delta H(\rho_w - \rho_i)/\rho_w$.
- (2) A removal of mass per unit area, δM , from the top surface causes the slab to rise a height $\delta M/\rho_w$.
- (3) In a similar way, ice melting from the bottom surface, or an addition of mass (snow) at the top surface, causes the slab to float lower.

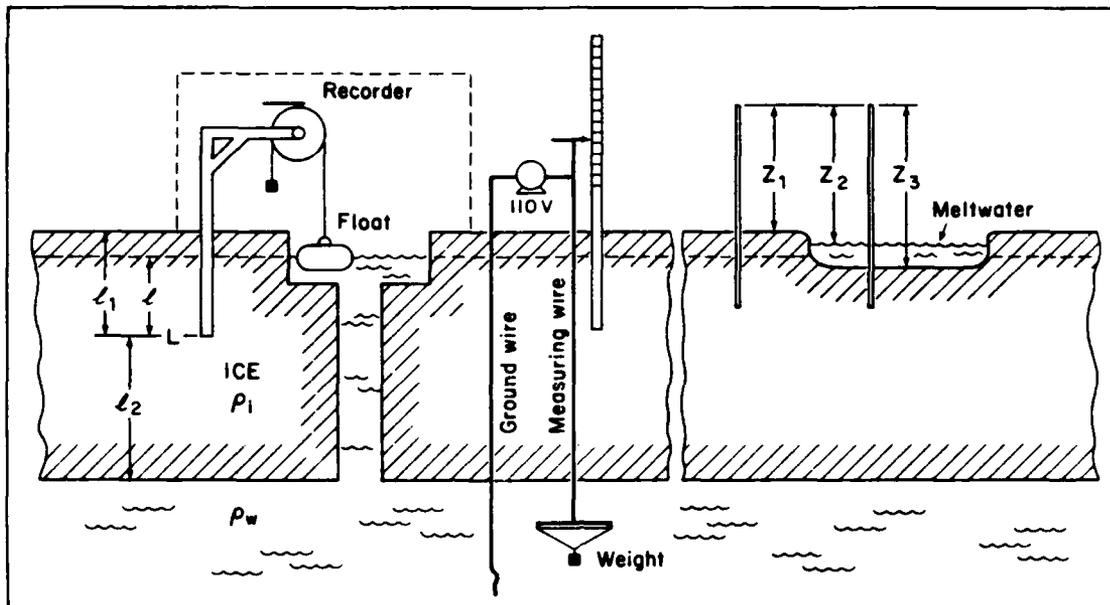


Figure 3. Original system of measuring sea ice mass balance developed by Untersteiner (1961). Thickness changes at the surface are measured by ablation/accumulation stakes, shown here in ice and meltwater puddles (lengths Z_1, Z_2, Z_3). Hydrostatic adjustments were recorded by a standard tide gauge. If we denote ρ_w as the density of seawater, changes of the freeboard level are related to ablation/accretion at the surface and bottom by $\Delta l = \Delta l_1 \rho_i / \rho_w - \Delta l_2 (\rho_w - \rho_i) / \rho_w$. The tide gauge was frozen into the ice at a fixed depth, L , and recorded changes of the freeboard level, Δl . With the measurements of the wire gauge, Δl_2 , and the stakes, Δl_1 , the equation is overdetermined, which provides a useful check on the composite accuracy. An intrinsic problem with the water level recorder is that it must be protected from wind and blowing snow by a suitable cover. This cover inevitably collects drifting snow, artificially exaggerating the readings of accumulation.

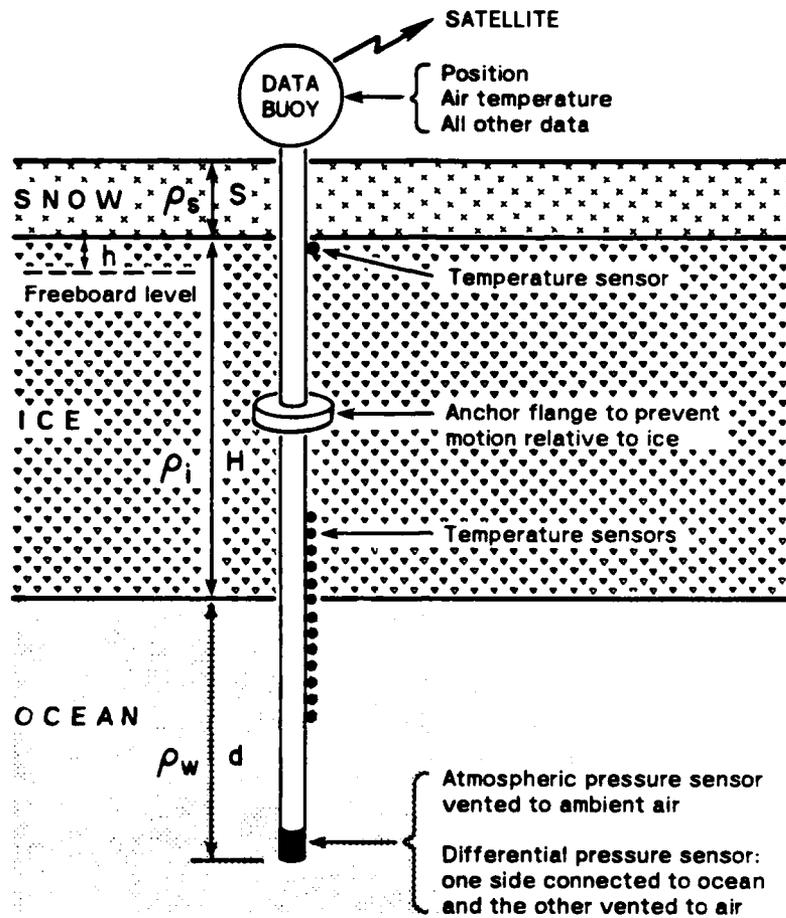


Figure 4. Schematic structure of the new data buoy for the measurement of ocean heat flux and ice mass balance. The temperature sensor at the ice-snow interface is added to obtain estimates of the temperature gradient across the snow layer (if any), to obtain estimates of heat conduction through the snow, and also to indicate the time when the snow layer becomes water-logged (0°C).

This can be expressed as

$$P \equiv P_{ocean} - P_{air} = g(\rho_w d + \rho_i H + \rho_s S), \quad (2)$$

where the terms d and S are defined in Fig. 4. By separating changes in ice thickness, H , into changes at the top surface, ΔH_t , and at the bottom surface, ΔH_b , where $\Delta H_b = -\Delta d$, we have

$$\Delta P/g + \Delta H_b(\rho_w - \rho_i) = \rho_i \Delta H_t + \rho_s \Delta S. \quad (3)$$

The terms on the left-hand side of Eq. (3) are to be measured, ΔP by the differential pressure sensor and ΔH_b by the thermistor string, as demonstrated in Fig. 2. The anticipated accuracy of the measurement is ± 0.5 cm for $\Delta P/\rho_w g$ and ± 1 cm for ΔH_b .

The right-hand side of Eq. (3) involves two unknown changes: melting of ice and its subsequent runoff from the top surface, ΔH_t , and changes in the mass of snow either by accumulation or by melting and runoff, $\rho_s \Delta S$. Except during a period in early summer when the ice surface consists of exposed ice and remnants of melting snow, the two processes at the right-hand side of Eq. (3) do not occur at the same time. They can be distinguished by the sign of ΔP . If ΔP is positive, snow accumulation accounts for the right-hand side ($\Delta H_t = 0, \Delta S > 0$). If ΔP is negative, mass is being lost from the top surface (initially $\Delta H_t = 0, \Delta S < 0$ and later in the summer $\Delta H_t < 0, S = 0$).

Over an annual cycle the snow runoff equals the snow accumulation (assuming equilibrium). This allows one to estimate the total ice loss, $\rho_i \Delta H_t$.

Changes in ice density are generally small and can be neglected for the present purpose. Changes in snow density are generally large, but are not important because only the snow mass contributes to the overall mass balance and only the product $\rho_s \Delta S$ appears in Eq. (3). Note also that only the mass that runs off the ice appears in the mass balance. Some of the meltwater stays at the surface in ponds and re-freezes (Untersteiner, 1961), but it affects neither the mass balance nor the differential pressure measurement.

Several additional sensors could be contained in the buoy to monitor environmental conditions not included in the mass balance equation but perhaps useful in other contexts. These include the atmospheric pressure, the air temperature, and the temperature close to the upper surface of the ice. As shown in Fig. 4, a separate sensor (vented to ambient air) for measuring atmospheric pressure is also located at the bottom end of the buoy tube to take advantage of the stable temperature environment prevailing in the ocean layer adjacent to the ice.

The measurements of pressure difference and temperature profile, when interpreted with the hydrostatic equation (3), lead to estimates of melting and freezing at the bottom surface and of snow accumulation and runoff from the top surface. Using the surface energy balance equation (1) gives an estimate of the oceanic heat flux.

A functional schematic of the proposed buoy, showing the connection between measured parameters, necessary calculations, and desired output data, is given in Fig. 5.

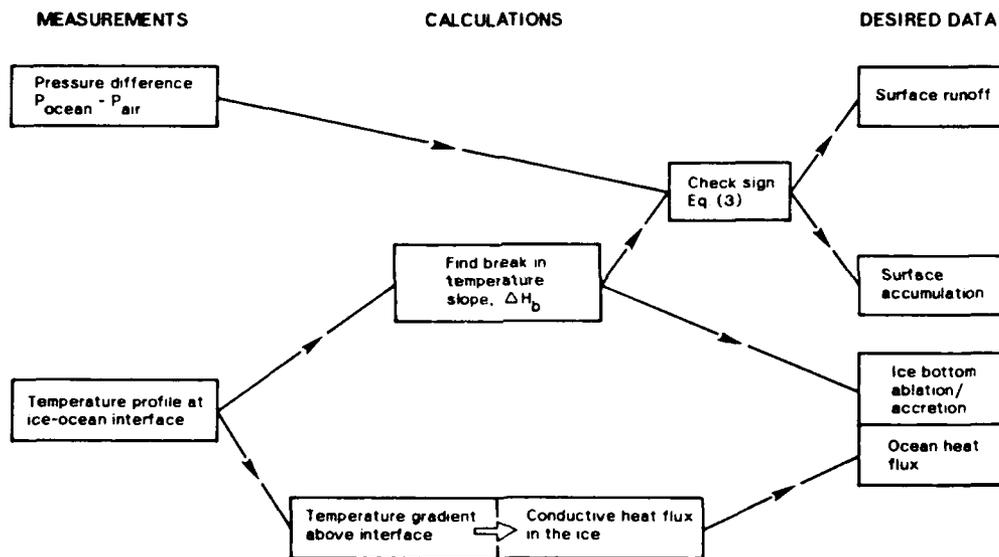


Figure 5. The measurements of pressures and temperatures are combined in this way to give estimates of the runoff or accumulation occurring at the upper surface, bottom melting or accretion, and the flux of heat from the ocean to the bottom surface of the ice.

4. SOME DESIGN FEATURES

As envisioned, the spar buoy hull will contain the pressure sensors at the lower end, thermistor circuits, an air temperature sensor, switching logic, an interface with the Argos transmitter, and a power supply. The Argos platform and the surface air temperature sensor would be those routinely used in the Arctic Data Buoy Program. The Argos platform is built by Polar Research Laboratories of Santa Barbara, California. Over 80 of these platforms have been used to date and have proved to be highly reliable (Untersteiner and Thorndike, 1982).

The absorption of short-wave, diffuse radiation (Grenfell, 1979) penetrating the ice surface tends, during the snow-free season, to loosen all objects inserted in the ice. To prevent this from happening to the pipe-shaped buoy hull, its surface must be coated with a substance that is highly reflective in the visible spectrum.

4.1 Limiting Circumstances

The buoy will initially float at its proper hydrostatic level. Upon freeze-up of the installation hole, the anchor flange will prevent vertical motion of the buoy relative to the ice. The new buoy is suitable for emplacement in ice much thicker than normal sea ice (i.e. in pressure ridges). The only limiting factor here is the technical means available in the field for drilling holes to the desired depth. (Systematic and long-term measurements at the keels of mature, solidified pressure ridges have not been made and would be particularly desirable.)

Another limiting factor in interpreting data from the new buoy is the temperature regime in late summer and autumn. Observations in 3-m thick ice in the Central Arctic (Fig. 6) show nearly isothermal conditions near the ice bottom at that time. The same is true of the adjacent layer of seawater. Maykut (1978, and personal communication) argues that the ocean heat flux is especially important during that season, when the location of the ice bottom will be particularly difficult to discern from the temperature profile.

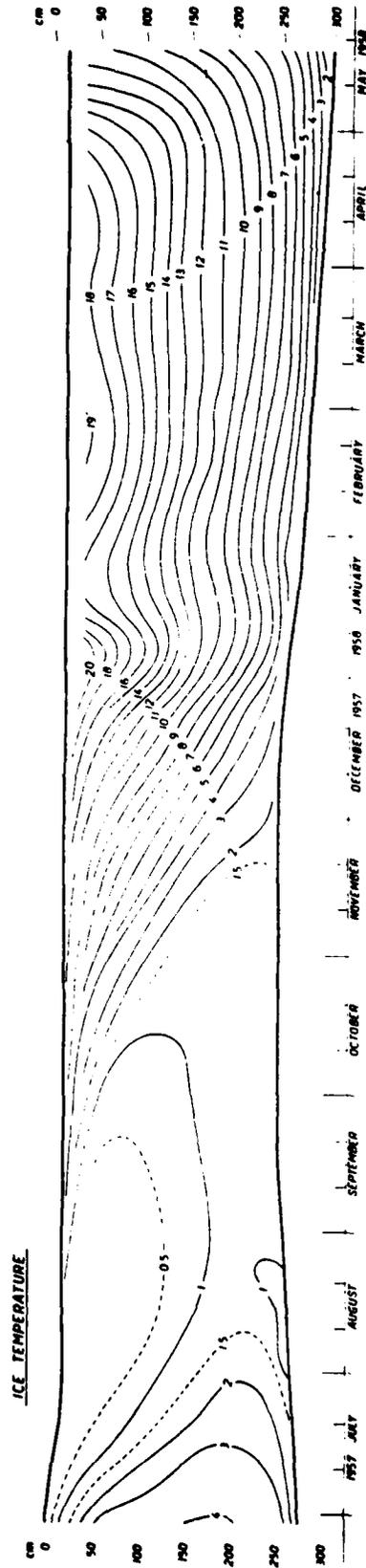


Figure 6. Isopleths of ice temperature in 3-m thick ice in the Central Arctic between July 1957 and May 1958. From August to November the conductive heat flux in the ice is nearly zero. (from Untersteiner, 1961)

However, the freshwater addition to the oceanic surface layer proceeds at a weather-dependent variable rate and frequently causes small-scale substrata in the uppermost ocean layer. Waves and eddies induced by the ice motion will cause small fluctuations in the water temperature while the ice temperature remains steady. This should be at least of some assistance in locating the base of the ice.

After November, the ice bottom can be positively located. Any hydrostatic depression of the ice during autumn is (at least largely) due to snowfall, so that by beginning of winter the new ice thickness and the total amount of ice lost at the surface and bottom should be known.

At some additional cost but no technical difficulty, two sensors could be added to the buoy:

- (1) An upward-looking, narrow-beam sonar transducer could be added to provide independent information on thickness changes occurring at the base of the ice. Such measurements have been made (Untersteiner, 1966), and the low number of observations needed would require an insignificant additional drain on the power stored in the buoy.
- (2) Random and accidental loading by pressure ice at the edge of the floe may induce a tilt in the attitude of the buoy. Within the whole set of errors in the other measurements, such a tilt would have to be very large to cause a significant additional, and unknown, error. Even so, a standard electrically indicating inclinometer in the buoy could solve that problem.

4.2 Field Testing

In 1983 dollars, the buoy is estimated to cost \$28,000, less than three times the cost of our expendable buoys that measure only surface pressure and location. If funding can be found, we would like to field test two prototype data buoys in a location where the ice velocity is not too high and where deployment and recovery can be accomplished easily, for instance, in the southern Beaufort Sea.

The two prototype buoys would be emplaced in ice of different thicknesses and with different snow depths. If the buoys are deployed in March and are operated through the subsequent summer, it should be possible to follow their functioning during all important phases (ice growth, snowmelt, runoff, and icemelt). As in the case of the Arctic Data Buoy Program, data from the new buoys would be obtained through Service Argos, France, by means of tapes periodically mailed to us. It is important to recover the buoys after the testing period so that they can be inspected for possible failure modes and design improvements, and for reuse.

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