THE INFLUENCE OF SNOWCOVER ON THE GROWTH OF ANTARCTIC FAST ICE

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Abstract

The physical properties of snowcover on Antarctic landfast sea ice have been observed to be highly variable spatially, temporally, and stratigraphically. This range of conditions is considerably greater than on Arctic sea ice; it is not adequately represented by the simple, homogeneous, approximations used in most thermodynamic ice growth models, which have been developed largely on the basis of observations made in Arctic regions. In this paper, measurements of the properties of the snowcover on the fast ice in McMurdo Sound are discussed along with their effects on ice growth. A simple thermodynamic model is described which can be used to simulate these effects over a wide range of ice, snowcover, and atmospheric conditions.

1. INTRODUCTION

The snowcover which develops on Arctic sea ice is characteristically thin, seasonal, and apart from a shallow depth hoar layer at the base, homogeneous. These properties have allowed the snowpack to be treated very simply in thermodynamic ice growth models without introducing significant error (Maykut and Untersteiner 1971, Semtner 1976, Gabison 1987, for example), despite the importance of snow to ocean/atmosphere energy transfers (Maykut 1978). In contrast, snowdepths on Antarctic sea ice can exhibit considerable spatial and temporal variations, primarily due to the large quantities of snow which blow off the Antarctic continent and are deposited along the coastal margins. Also, in the absence of surface melting during the summer, the snow on second and multi-year ice can develop complex stratigraphies through the deposition of seasonal layers. This variability cannot be adequately represented by the simple one layer, homogeneous, approximations used in most thermodynamic models. In particular, extremely deep, near-shore snowpacks frequently cause surface slushing and snow-ice formation; this produces a three-layer system of conductors (snow, snow-ice, and ice), all of which must be accounted for if accurate ice growth predictions are to be achieved. Observations of snow accumulation and fast ice growth in McMurdo Sound, Antarctica, have led to the development of a simple thermodynamic model which incorporates these phenomena, and can be used to investigate the effects of a wide range of snow conditions on thermodynamics and ice growth.

2. SNOWCOVER PROPERTIES

During the austral winter of 1986 a project was undertaken to measure the dynamics, thermodynamics, and physical properties of the fast ice cover in McMurdo Sound, Antarctica, (Crocker 1988). In addition to the ice observations, regular measurements of the properties of the snowcover on the ice were made at a site roughly 1.5 km directly offshore from Scott Base (see figure 1). The fast ice under study was three years old, and at the beginning of the measurement period (early April) was covered by a homogeneous, dense, granular, snowcover 0.3 m in depth. Within a month, a new layer of felt-like snow had been deposited

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and a thin but distinct depth hoar layer had developed between the new and old layers. This general structure remained unchanged throughout the remainder of the winter, although the old snow layer slowly compacted and increased in density over the course of the season. This change can be seen in table I, which shows the thickness (h_s) and density (ρ_s) of the old and new layers in early (May) and late winter (September).

TABLE I. Snowcover Characteristics

		May 19	September 26
Old	h_s (m)	0.29	0.12
	ρ_s (kg m^{-3})	405	522
New	h_s (m)	0.20	0.46
	ρ_s (kg m^{-3})	345	340

The thermal conductivity of the snow was calculated using the empirical formula of Abel's (1893)

$$k_s = 2.85 \times 10^{-6} \rho_s^2 \tag{1}$$

where k_s is in W m^{-1} °C^{-1} and ρ_s is in kg m^{-3}. The presence of three distinct layers in the snowpack, however, introduces a significant error in the thermal conductivity estimate if a bulk density is used. To overcome this, the three layers (new snow, hoar, old snow) are treated as series connected conductors. If the density of the hoar layer (which was too thin and fragile to be sampled using the available equipment) is fixed at 150 kg m^{-3}, then the mean thermal conductivity of the three layered snowpack is roughly 1.4 times less than that calculated using a bulk density. For modelling purposes, k_s is held constant, as the observed range of snow conditions produced only slight variations in the calculated thermal conductivity. Changes in snow depth are simulated by a linear increase from 0.30 m on April 1 to 0.65 m on May 1.

3. MODEL DESCRIPTION

A complete description of the model which will be used can be found in Crocker and Wadhams (1989), and only the basic formulae will be given here. It is a modified version of the thermodynamic model developed for climatic studies by Semtner (1976), and, like most such models, utilizes a calculated equilibrium surface temperature (T_0) from which all the relevant fluxes, and hence ice growth, can be determined. This is achieved by setting the flux of energy from the surface to the atmosphere (F_a) equal to (but with opposite sign) the flux of energy through the snow/ice slab (F_0), that is

$$F_a + F_0 = 0. \tag{2}$$

The flux to the atmosphere is

$$F_a = F_s + F_l + F_{Ld} + (1 - \alpha)F_r + \varepsilon\sigma T_0^4 \tag{3}$$

where α is the surface albedo, ε is the surface emissivity, σ is the Stefan-Boltzmann constant, and the subscripts s, l, Ld, and r, refer to the sensible heat flux, latent heat flux, incoming long-wave radiation, and incoming short-wave radiation respectively. The conductive flux through the slab is calculated using

$$F_0 = \frac{(T_b - T_0)}{h_i/k_i + h_{si}/k_{si} + h_s/k_s} \tag{4}$$

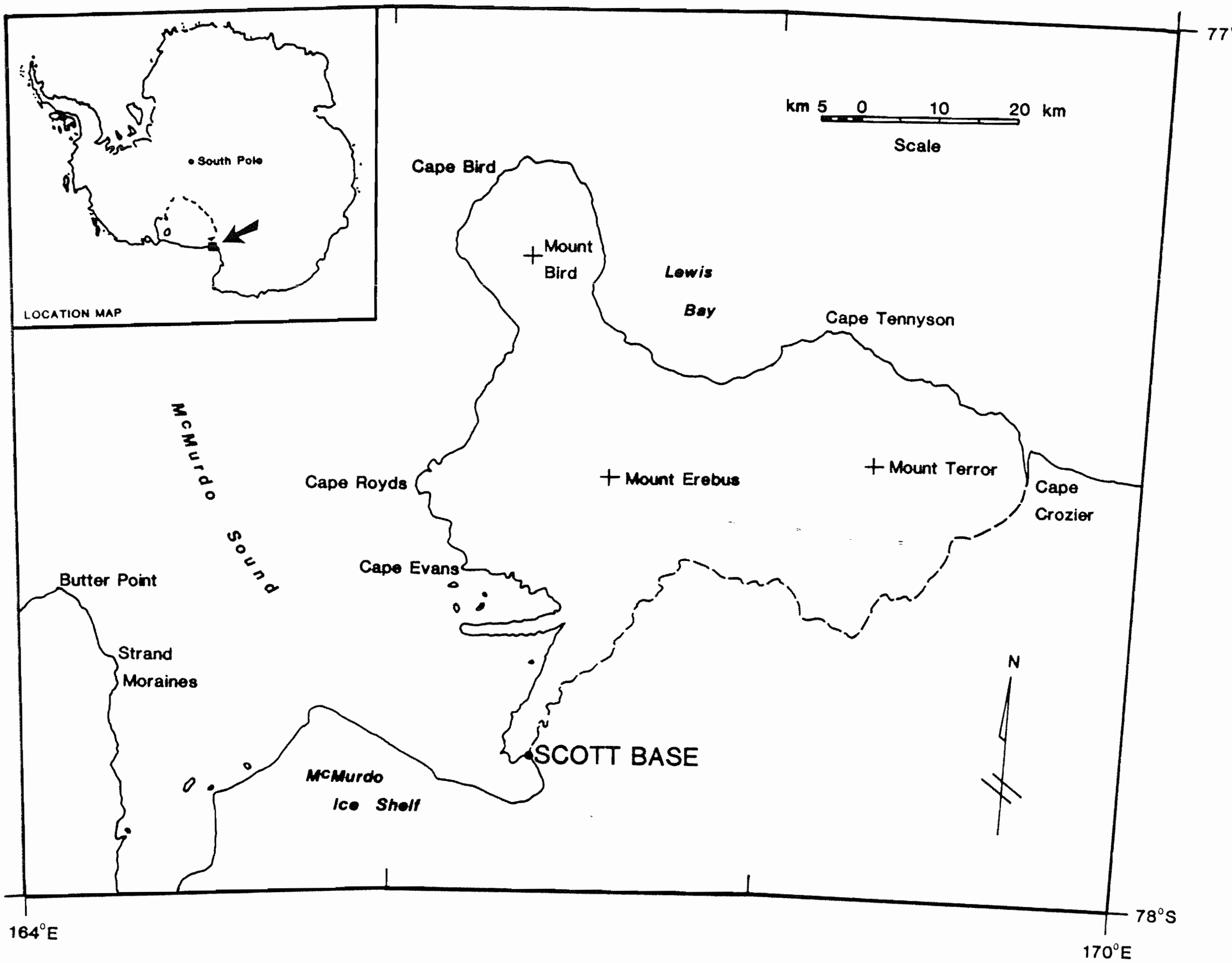

Figure 1: Location map showing the McMurdo Sound Region. The main study site was 1.5 km offshore from Scott Base.

where T_b is the temperature at the ice/water interface (−1.8 °C) and the subscripts i, si, and s correspond to ice, snow-ice, and snow respectively. Fluxes of sensible and latent heat are determined from

$$F_s = \rho_a c_p C_s u(T_a - T_0) \tag{5}$$

and

$$F_l = 0.622 \rho_a C_l u(f e_{sa} - e_{s0})/p_0, \tag{6}$$

where ρ_a is the density of air, c_p is the specific heat of air, C_s and C_l are the energy exchange coefficients, u is the mean wind velocity, T_a is the mean air temperature, f is the relative humidity, p_0 is surface pressure, and e_{sa} and e_{s0} are the saturation vapour pressures in the air and at the surface. The saturation vapour pressures are calculated from the air and surface temperatures using the fourth order polynomial approximations of Maykut (1978); F_{Ld}, F_r, u, and T_a are modelled using Fourier transforms of mean monthly data from a variety of sources, and input explicitly. Substitution of equations 3 through 6 into equation 2, allows T_0 to be determined by a Newton-Raphson or similar iterative process. Once the equilibrium surface temperature is known, F_a and F_0 can be found and used to calculate ablation or accretion at the base and surface of the slab. Basal melting or growth is then

$$\Delta h_b = \Delta t(F_0 - F_b)/q_b, \tag{7}$$

where Δt is the time increment, and q_b is the volumetric heat of fusion at the ice/water interface. For snow covered ice, surface melting is calculated by

$$\Delta h_s = \Delta t(F_a - F_0)/q_b \tag{8}$$

and for bare ice, surface ablation is found using

$$\Delta h_i = \Delta t(F_a - F_0)/q_i, \tag{9}$$

where q_s and q_i are the volumetric heats of fusion for the snow and ice respectively (note that since the latent heat of fusion of sea ice is salinity dependant, q_b and q_i, are not necessarily the same). Surface slushing and snow-ice formation are assumed to occur as soon as two criteria are met; a positive hydrostatic water level, and a surface temperature in excess of −8 °C. The later is somewhat speculative and represents a temperature at which interconnecting channels in the ice permit throughflow. This particular value has been chosen on the basis of observations in McMurdo Sound, where the snow load over a section of the ice sheet was intentionally increased, and the ice/snow interface temperature carefully monitored. A more rigorous approach to the problem should include the effects of salinity and crystal structure on the permeability of the ice, but current theory and observations are inadequate for such a analysis. The hydrostatic water level (H_w), on the other hand, is relatively simple to compute if reasonable estimates of the snow and ice densities are available. After Röthlisberger (1983),

$$H_w = \frac{\rho_i}{\rho_w} h_s - \frac{\rho_i}{\rho_s} h_i \left(1 - \frac{\rho_i}{\rho_w}\right). \tag{10}$$

The thermal conductivity of snow-ice (k_{si}) has been set to $0.9 \times k_i$, where k_i is

$$k_i = k_0 + \frac{\beta S}{T}, \tag{11}$$

k_0 is the thermal conductivity of pure ice (2.03 W m^{-1} °C^{-1}), S is the ice salinity in parts per thousand, β is a constant (0.117 W m^2 °C^{-1}), and T is the ice temperature in degrees Celsius (Maykut and Untersteiner 1971). This formula for k_{si} was arrived at through an analysis of the thermal conductivity of saline, bubbly, ice using the theory of Schwerdtfeger (1963), and reflects the influence of the high salinity and high air volume fraction. Also, it is assumed that flooding has no effect on the thermodynamics of the system (apart from the obvious effects of changing h_s and h_{si}), since the flooding liquid is highly saline brine with a temperature very close to the snow/ice interface temperature, and the temperature gradient through the slab is largely unaffected. Additional changes to the original Semtner model have been made in order to incorporate the extremely large negative oceanic heat fluxes and associated sub-ice platelet formation, which characterize the McMurdo Sound region, and to account for their effect on ice growth. In this form, the model produces ice growth predictions which have been shown to be in good agreement with the McMurdo Sound observations (Crocker and Wadhams 1989), and can therefore be used to investigate the effects of the observed range of snow conditions on both seasonal and multi-year cycles of fast ice growth and decay.

4. RESULTS AND DISCUSSION

4.1 Annual Growth Cycles

Predicted changes in snow-ice and columnar ice thicknesses produced by different snow accumulations are shown, over a one year period, in figure 2. The complete absence of snow-ice with peak snow depths of 0.6 m or less is immediately evident, as is the constancy of h_{si}, both temporally and with different snow accumulations. Slushing occurs only after enough snow has been deposited to provide sufficient thermal insulation for the ice surface to warm to the critical temperature of −8 °C. This occurs at roughly the same time in each case, although it is slightly earlier for the larger accumulation rates. The depth of this initial flooding is quite large because H_w is allowed to build up as h_s increases and slushing is inhibited by low ice surface temperatures. It is similar for all snow accumulation rates (for $h_s > 0.6$ m) because the snow depth required for sufficient surface warming is the same, and therefore H_w at the time of first flooding is the same. Hydrostatic water levels remain positive throughout the winter but the surface temperature (which is reduced after flooding effectively reduces the thickness of the low conductivity snow layer) is too low to allow the fluid to flow through the ice. This condition was observed to exist in McMurdo Sound throughout most of the winter. For extremely large accumulations, however, or when accumulations occur over a longer period of time, such as in areas immediately adjacent to the shore, several slushing events may take place over the course of the season. The slight increase in h_{si} at the end of the year, with a snow accumulation of 1.0 m, is the result of increasing air temperatures and decreasing columnar ice thicknesses. This allows greater heat flow to the ice surface from both above and below, and increases T_s to the critical value. This is augmented by increased hydrostatic water levels as the columnar ice thins, and has also been observed in McMurdo Sound in spring. The peak columnar ice thickness decreased significantly with increasing snow accumulation down to 0.6 m a^{-1}, where there is a marked increase, followed by a more gradual decline. This is caused by the formation of the snow-ice layer which reduces the thermal resistivity of the slab by reducing the snow depth. The subsequent reductions in columnar ice growth again reflect the increased insulation provided by the deeper snowcovers remaining after slushing.

4.2 Multi-Year Growth Cycles

The model can also be used to investigate the growth characteristics of the ice over several seasonal cycles. Although multi-year ice is not common in the Antarctic, fast ice in secluded bays and inlets can reach considerable ages and thicknesses (Hoffman and Stehle 1963). In fact, the ice in the southern end of McMurdo Sound, on which the measurements were made, was three years old at the start of the sampling period, and much older and thicker sections were present between the southern side of Pram Point and the McMurdo Ice Shelf. Figure 3 shows five annual cycles of growth and decay for McMurdo Sound fast

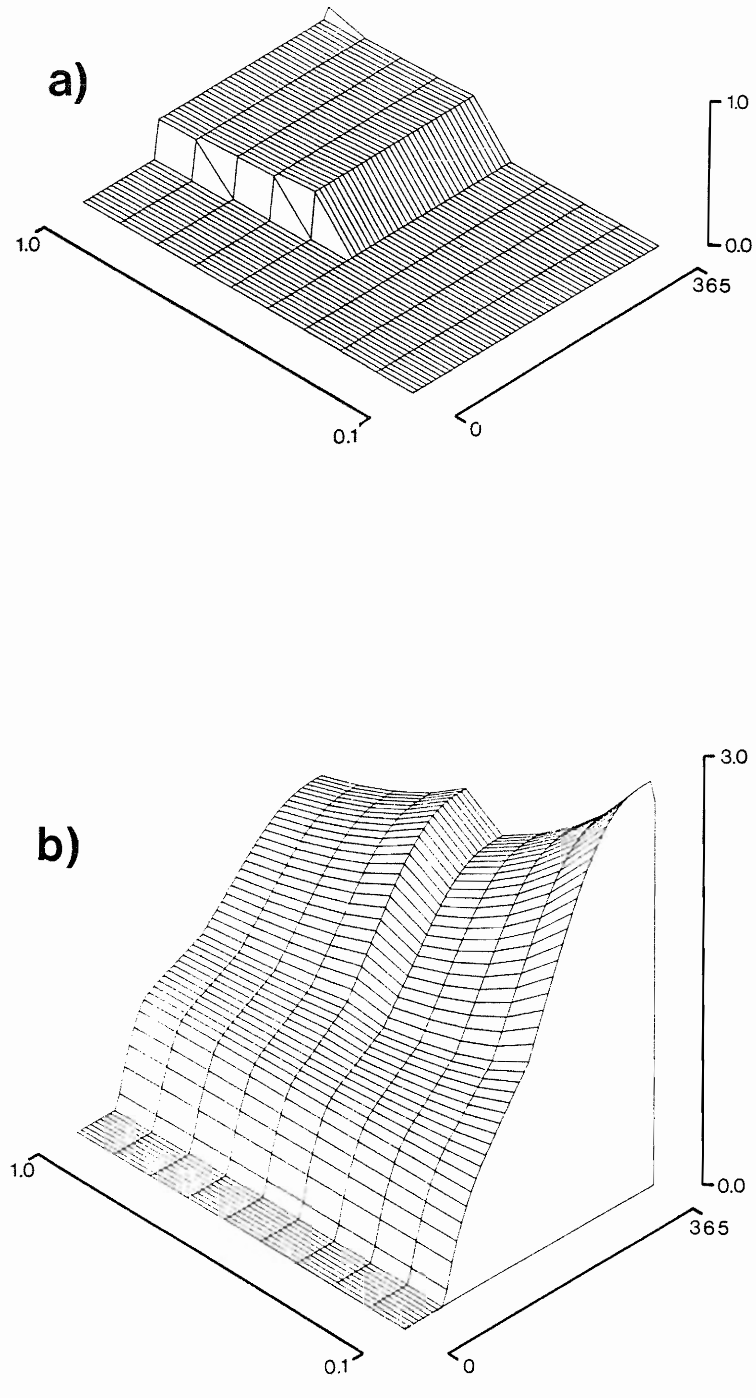

Figure 2: Predicted snow-ice (a), and columnar ice (b), thicknesses for McMurdo Sound with a range of annual snow accumulations. The x, y, and z axes are; time (Julian days), accumulation rate (m a^{-1}), and thickness (m), respectively.

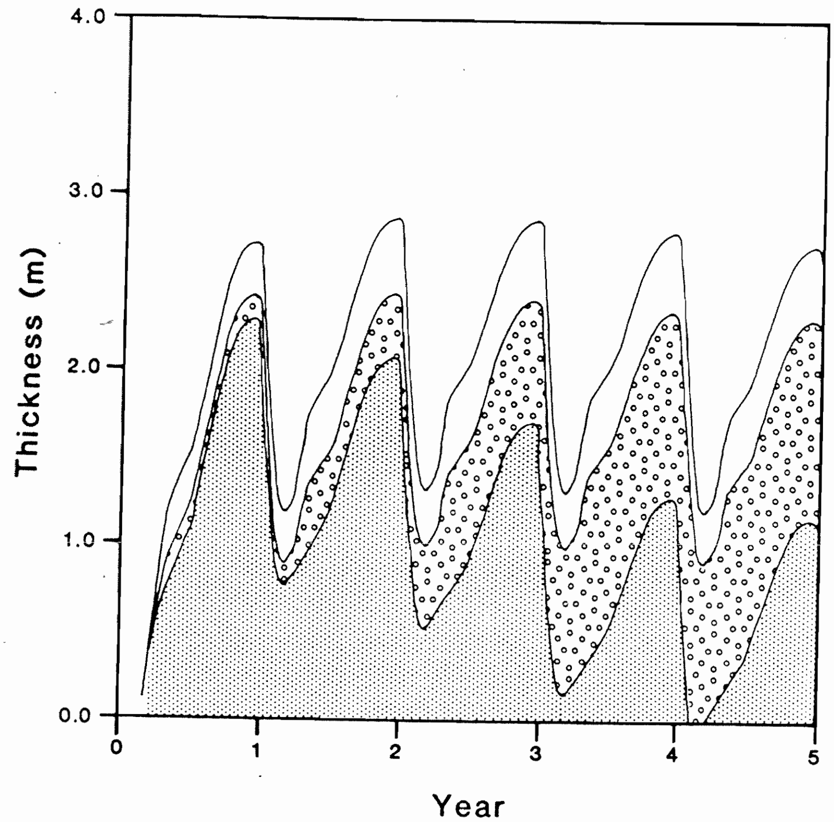

Figure 3: Five predicted annual growth and decay cycles for McMurdo Sound with the observed snow accumulation rate of 0.35 m a^{-1} (columnar ice is represented by the shading, snow-ice by the open circles, and snow by the white area under the upper line.)

ice using the observed snow accumulation rate of 0.35 m a^{-1}. The model predicts that snow-ice will not form in appreciable quantities until the second summer, but that by the fifth year equilibrium conditions will be established, with snow-ice constituting just under 40% of the 1.7 m thick ice sheet at peak spring thickness. It is interesting to see that the total ice thickness decreases steadily after the first winter, in direct contrast to multi-year ice in the Arctic. This is due to the large positive oceanic heat fluxes and basal ablation during the summer, and the reduced winter growth associated with the deepening snowpack, which is not removed by surface melting. Although insufficient field observations are available to confirm such a decrease, measurements of extremely thin (< 1.0 m) multi-year ice support the general conclusion that old ice does not necessarily grow to great thicknesses. The equilibrium conditions, and the number of years taken to reach them, are strongly affected by the snow accumulation rates. Increasing the annual snow deposition from 0.1 m to 0.6 m reduces the maximum thickness of the columnar ice by about 30%, and surprisingly causes h_{si} to drop by almost 35%. This is product of the reduced columnar ice growth associated with the deeper snowpack, which is completely removed by basal melt during the summer and leaves the snow-ice exposed to the warm ocean. The time required for equilibrium conditions to be established decreases from 11 years with 0.1 m a^{-1}, to 4 years with the deposition of 0.6 m of snow per year. At accumulation rates above 0.6 m a^{-1} the pattern becomes complex, and true equilibrium conditions are not produced. Instead a repeating 3 year cycle is established, with large annual variations in h_i and with much greater snow-ice thicknesses.

5.CONCLUSIONS

The physical properties of the snowcover on Antarctic fast ice are highly variable and strongly influence the development of the ice sheet. The wide range of possible snow depths, and the formation of snow-ice under extremely deep snowpacks, are particularly important; they have been incorporated into a thermodynamic model which can be used to investigate their effects on ice growth. Despite the overall simplicity of the model, and the rather crude treatment of the conditions required for slushing and snow-ice formation, it is shown to produce results in good agreement with the limited observations. Although more rigorous testing over a complete range of climatic conditions will be possible only when more detailed measurements of the crystal structure of Antarctic fast ice, and the temporal variations in its snowcover properties become available, the analyses clearly demonstrate the importance of snow to ice sheet development, and the need for better parameterization of the snow in this type of model.

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