

ILLUSTRATIONS OF BIOLOGICAL ROLES OF WINTER LAKE COVER

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ABSTRACT

The major components of the winter cover of lakes in eastern North America are described within the context of the processes which control their evolution. Examples of the distinctive biological roles of the components and processes are then discussed including their roles in the energy budget, in nutrient loading and in the light and oxygen regimes of the lake. Mention is also made of the ways in which the various components promote or mute the effects of ice-shove on the lake margins.

INTRODUCTION

The principal components of the typical winter cover of lakes in eastern North America are portrayed in part C of Figure 1. The 'black ice' component comprises lake water, frozen in situ into a sheet of vertically-oriented, columnar, crystals. Above the black ice is the granular 'white ice' component formed by the freezing of slush produced when the snowcover is flooded as a result of cracking while the ice sheet is depressed below the hydrostatic water level (HWL). Figure 1 is designed to portray the process of white ice formation, the sequence shown may occur several times during a winter. The third cover component is a layer of snow which may develop quite distinctive characteristics in part as a result of periodic floodings and incorporation of its basal layers into the white ice (Adams and Prowse, 1978). The processes of evolution of lake winter cover and the forms associated with them are discussed in some detail in Adams 1976a and b.

The thicknesses of the various components and their relative proportions may vary greatly within a single lake, between lakes in the same region, between regions and between winters. Also, naturally, they vary greatly at a single point during a winter. Data from the Kawartha Lakes near Peterborough, Ontario, and from Lake St. George, 32 km north of Toronto, Ontario, give an indication of the order of magnitude involved for one particular season (Fig. 2). In particular situations, one or two of the components may be entirely absent. The case of Coon Lake (Fig. 3) in 1979-80, a very low snowfall year, illustrates this point.

The purpose of this paper is to identify particular biological roles of the cover components and of the winter cover as a whole with a view to drawing attention to the importance, in biological limnology, of an awareness of the true characteristics of the cover of study lakes. Examples chosen are drawn from studies completed or in progress at Trent University, Peterborough, Ontario.

Winter Cover and Oxygen Regime

During the ice-free season, a lake is directly affected by the turbulent overlying atmosphere. There is a diffusion of oxygen across the air/water interface and the lake water is oxygenated as a result of wind-induced stirring as well as by the thermally induced turbulence of spring and/or fall turnovers. An ice cover cuts the lake off from the direct influence of atmosphere so that, where flushing rates are low, the unfrozen

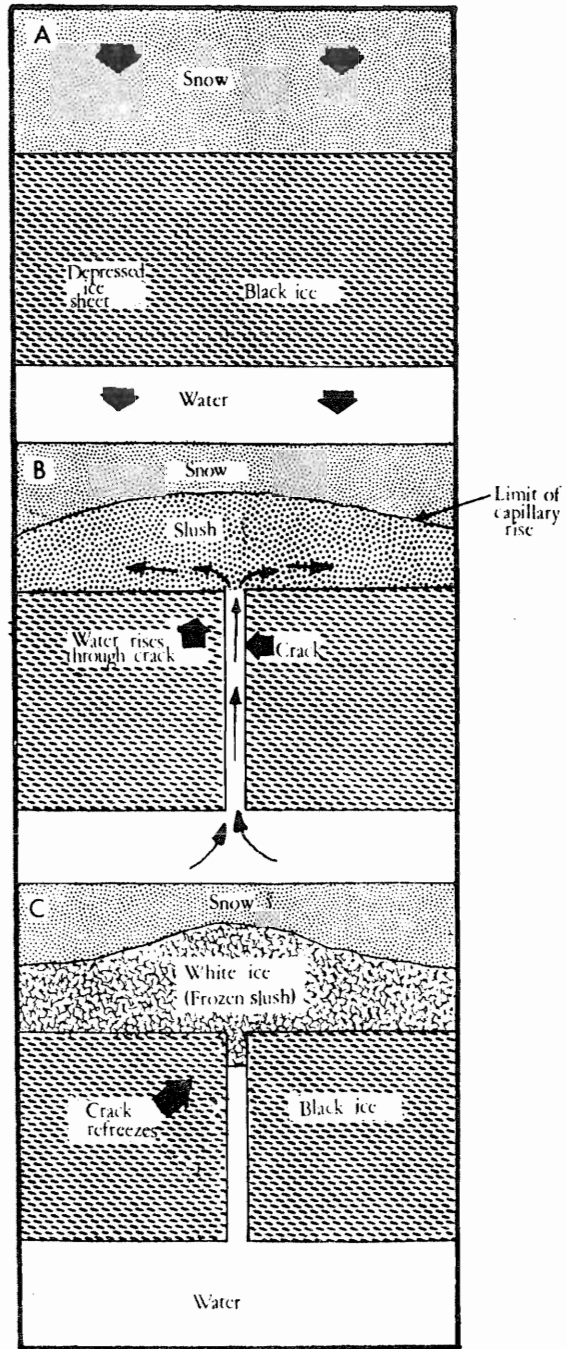


Figure 1. The formation of white ice from slushed snow.

water body is characterized by a 'reversed' (cold at the top, warm at the bottom) thermal stratification which is stable despite the fact that only a small (0-4°C) range of temperature is involved.

In this 'sealed off' situation, oxygen continues to be consumed through biological activity and by the breakdown and oxidation of materials in the water. This consumption of oxygen is referred to as Water Oxygen Demand (WOD) in the limnological literature (e.g., Hutchinson 1957, Wetzel 1975). The biological component of this represents the respiration of organisms in the water, fish, zooplankton, bacteria, etc. Also, oxygen is consumed through oxidation and biological activity in the layers of sediment which underly the lake system. This is referred to as Sediment Oxygen Demand (SOD).

The lake system does not rely entirely on direct contact with the atmosphere for its oxygen supply. Given light, oxygen is produced by photosynthesis (primary production) within the lake. Rates of primary production are controlled by the light regime which, in winter, is controlled to a greater or lesser extent by the snow and ice cover of the lake.

In addition to limiting atmosphere/water contact and, through its effect on light, affecting photosynthesis, the ice cover of a lake has two other important effects on its oxygen regime. As water freezes, there is a more or less efficient exsolution ("freeze-out", Pounder 1965) of dissolved solids and gases including oxygen. In the case of black ice, which essentially thickens downwards into the lake, this implies a contribution of oxygen to the unfrozen water body from its frozen surface layers. This is a gain to the unfrozen part of the lake, not a gain to the lake as a whole (defined so as to include the ice cover). However, the slushing process portrayed in Figure 1 represents a loss to the water body (and possibly to the whole system) of water and oxygen from immediately below the ice.

The oxygen regime of lakes in southern Ontario has been extensively studied by Dr. D.C. Lasenby of Trent University and his students, notably M.B. Jackson (see also the follow-up of Jackson's work by Winter, 1980).

Table 1

Morphological Characteristics of Lakes Cited (see Jackson, 1980)

Lake	Lake Area (A) (m ² x 10 ⁴)	Volume (m ³ x10 ⁴)	Mean Depth (m)	Maximum Depth (m)	Catchment Area (Ac) (m ² x10 ⁴)	Ac/ A
St. George ¹	10.3	57.9	5.6	16	132	12.8
Coon ²	53.1	213.3	4.0	18	239	4.5
Big Cedar ³	224	1411	6.3	18	1245	5.6

1. 43°45'N, 79°30'W, 296 m above sea level (a.s.l.)

2. 44°36'N, 78°12'W, 258 m a.s.l.

3. 44°36'N, 78°11'W, 250 m a.s.l.

Jackson (1979 and 1980) made a detailed study of the winter oxygen regime of the unfrozen water in three lakes, two of them in the Kawartha lakes, (see Table 1), using the relationship:-

$$\text{Total Oxygen Change} = (P + F) - (\text{WOD} + \text{SOD} + S) \quad (1)$$

Where: P = oxygen contributed by primary production
 F = oxygen contributed by black ice freeze out
 WOD = oxygen consumed in the water column (as described above)

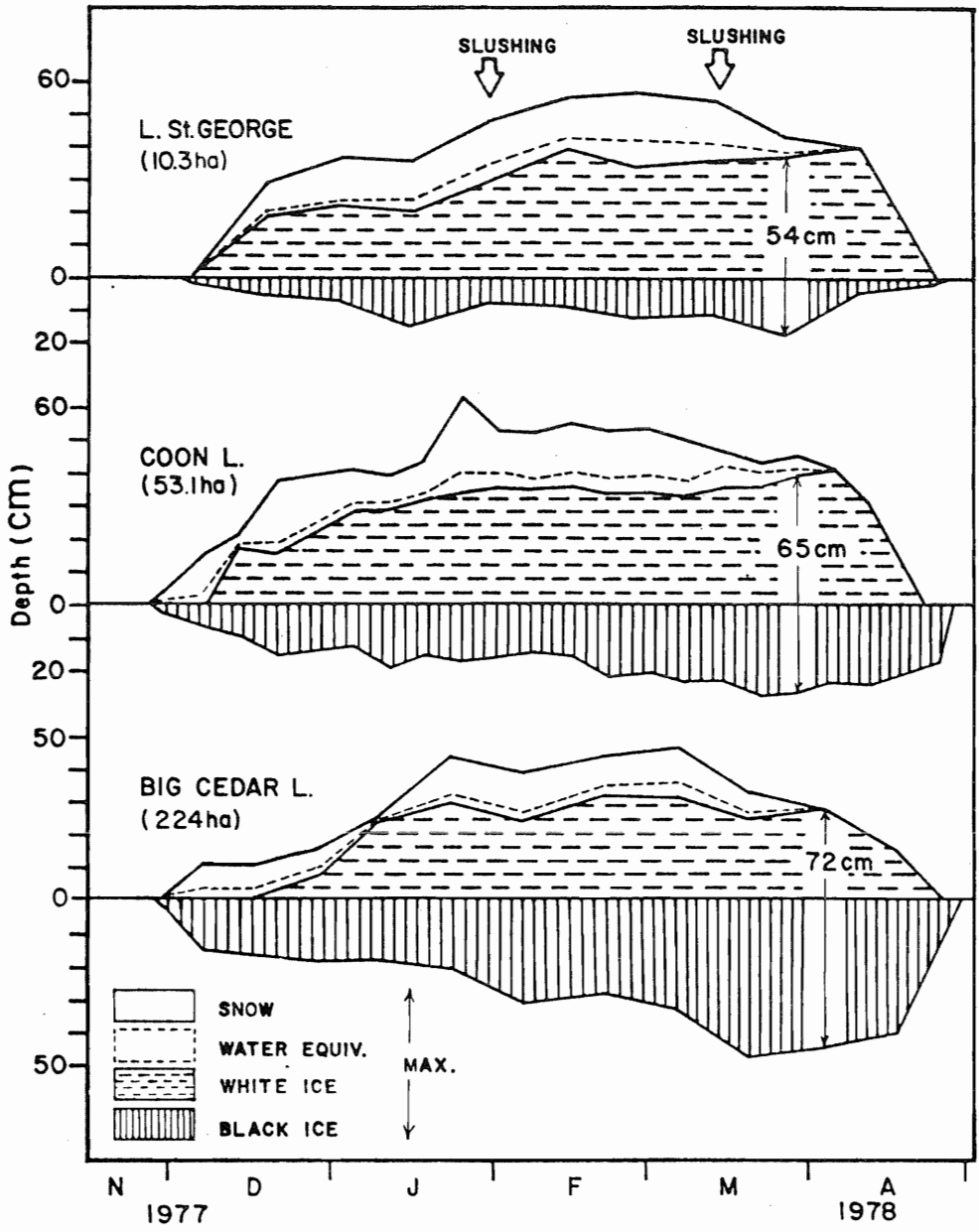


Figure 2. Snow and ice characteristics at maximum water-depth stations in the study lakes, Nov. 29/77 - Apr. 26/78.

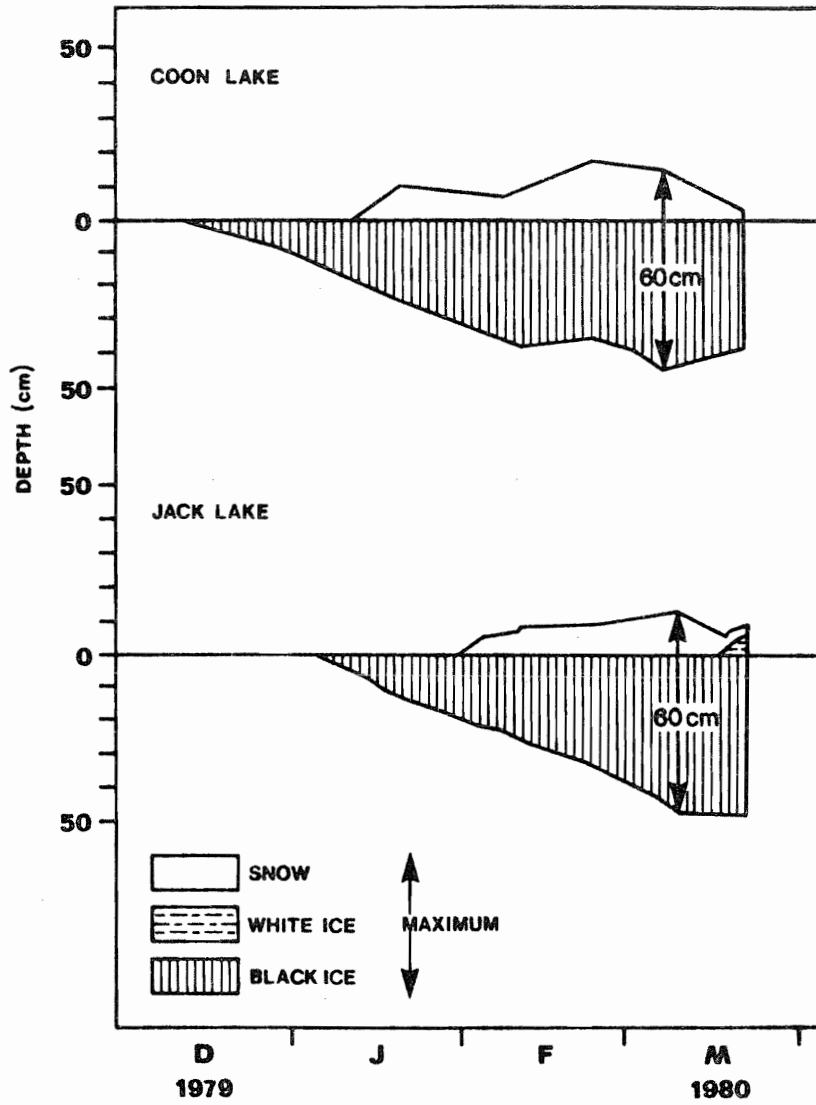


Figure 3. Evolution of ice cover on two Kawartha Lakes, 1979-80 (after Winter, 1980).

SOD = oxygen consumed in the sediment layer (as described above)
 and S = oxygen removed from the water body as a result of slushing process (also described above)

This equation assumes either negligible throughput of water during the winter or a balance between oxygen gains resulting from inflows and oxygen losses due to outflows. Jackson's method of calculating the quantities of oxygen involved in S and F, which are interesting here as they involve knowledge of the processes and forms of the lake cover, is presented in Table 2. His methods for calculating the other terms can be obtained from the references cited. The winter covers concerned were studied in detail by Prowse, (1978), and their general development is portrayed in Figure 2.

Table 2

Calculation of the F and S Terms of Equation 1

A. Calculation of F = oxygen contributed by freeze-out* (mgO₂/cm²/day) = $V_b \cdot [O_2]_o \cdot \rho_b / A \cdot t \cdot 10$

Where: V_b = volume of black ice at maximum ice thickness (m³)
 [O₂]_o^b = average oxygen concentration at 0m (mgO₂/l)
 ρ_b = density of black ice (0.9 g.cm⁻³, see Adams and Lasenby 1978)
 A = lake surface area (m²)
 t = time period of whole-lake oxygen loss (days)

* 100% freeze-out is assumed

Ice thicknesses were obtained from measurements in at least two drill holes per lake. Ideally, for calculations of this sort, such measurements would encompass the full range of values on a lake. Note that this calculation requires the thickness of black ice not the total thickness of the ice cover. The assumption of 100% freeze-out gives a high value as, as bubble layers in black ice clearly testify, exsolution must rarely be 100% efficient.

B. Calculation of S = oxygen consumed by slushing (mgO₂/cm²/day) = $V_w \cdot [O_2]_o \cdot \rho_w / A \cdot t \cdot 10$

Where: V_w = volume of white ice at maximum ice thickness (m³)
 [O₂]_o^w = average oxygen concentration at 0m (mgO₂/l)
 ρ_w = density of white ice (0.89 g.cm⁻³ but see Adams and Lasenby 1978)
 A = lake surface area (m²)
 t = time period of whole-lake oxygen loss (days)

The loss of oxygen from the unfrozen water body comprises only that proportion of the white ice which is formed of lake water. The remainder of the white ice is formed of snow partly melted and then refrozen. An accurate calculation of this term therefore requires a knowledge of the density of snowpacks flooded to form white ice. The oxygen represented by the snow portion of the white ice has to be subtracted from S as calculated above to obtain the oxygen budget term for white ice.

Jackson's 1977-78 oxygen budgets for three lakes are presented in Table 3. The oxygen budget for Big Cedar Lake is presented, in percentage terms, as an example, in Equation 2.

$$\text{Total Oxygen change} = (10 + 13) - (33 + 80 + 10) \quad (2)$$

From the point of view of those interested in the lake cover, it is interesting to note that while the principal oxygen losses were associated with WOD and SOD in each case, with the latter tending to dominate, slushing (S) accounted for between 2 and 10% of the total. By comparison, consumption by fish respiration (<0.2%) was negligible.

Table 3

Contribution to Whole-lake Oxygen Change from SOD, WOD, Fish Respiration, Slushing,
Freeze-out and Primary Production in the Study Lakes, Dec. 5/77-Apr. 19/78

Oxygen	L. St. George		Coon L.		Big Cedar L.	
	Rate (mgO ₂ /cm ² /day)	%	Rate (mgO ₂ /cm ² /day)	%	Rate (mgO ₂ /cm ² /day)	%
Loss						
SOD	0.010	23	0.013	45	0.024	80
WOD	0.024	55	0.010	34	0.010	33
Fish Respiration	0.000	<0.2	0.000	<0.2	0.000	<0.2
Slushing	0.001	2	0.002	7	0.003	10
Freeze-out	0.001	2	0.001	3	0.004	13
Primary Production	0.003	7	0.003	10	0.003	10
Total						
Estimated	0.031	70	0.021	72	0.030	100
Observed	0.044	100	0.029	100	0.030	100
Difference	0.013	-30	0.008	-28	0	0

Gains of oxygen from the two sources, primary production (P) and freeze-out (F) were of similar magnitude, 7-10% and 2-10% respectively.

Naturally, the magnitude and proportions of these terms must vary greatly. A shallow lake (a relatively small unfrozen water body) in a snowy winter, for example, might experience relatively high losses through slushing in conjunction with small gains from freeze-out (little black ice) and from primary production (photosynthesis limited by low light conditions). This is a combination of conditions in which fish kills would be likely. By contrast, a year of low snowfall, such as the 1979-80 winter in southern Ontario (portrayed in Figure 3), will produce relatively high gains from freeze-out (thick black ice) and from photosynthesis (high light levels, see section on light below), with small or no losses from slushing. This is well documented for Coon Lake, in the Kawarthas, by Winter 1980 (p.25f). This, with the complication of the period of 24 hours of darkness, is the normal condition in the high arctic. Various combinations of lake morphology and cover characteristics can be envisaged.

With regard to oxygen regime, then, it is important not to view the winter cover of a lake as a lid or seal which is more or less inert. Rather than being static, the cover is a dynamic phenomenon resulting from the interaction of "the atmosphere" and "the lake". The processes of cover formation continue, albeit in a much modified fashion, water-atmosphere exchanges even to the point where unfrozen water is brought directly into contact with the free atmosphere. It would be interesting to pursue, at a more or less micro scale, the roles in the winter oxygen regime of dry cracks in an ice sheet and of wet cracks during the periods when slush layers are freezing to form white ice. The roles of such wet cracks is touched on again below.

Winter Cover and Nutrient Loading

Although the water body of an ice covered lake is effectively shielded from atmospheric turbulence and the mass and energy exchanges associated with it, it is affected by precipitation receipts although in ways which are very different from those of the ice-free season. Except in situations where rain and/or snow meltwater enter the unfrozen water body through cracks in the ice, winter precipitation tends to remain on or in the ice cover until break-up. The precipitation concerned is a gain in mass to the system in the sense that the water level is raised after each precipitation event. Also, it is very much a part of the system in the sense that it forms a key component of the winter cover which has wide ramifications for its physical and biological aspects. However, such precipitation generally remains effectively apart from the unfrozen water body. In the spring, usually after snow has melted on land, snow on the ice melts and percolates into the lake. After the lake snowcover has disintegrated, the white ice cover melts contributing its share of winter precipitation to the water body.

Lake levels rise in the spring because of inputs of meltwater from the land but precipitation which fell directly onto the lake cover does not, on melting, add further to the hydrostatic water level. However, it contributes anew to the system in the sense that it increases the available supply of liquid water and in the sense that its constituents become available to the water body. These same points can be made for the spring melt contribution of ice, such as black ice, which is not precipitation-derived.

The "constituents" of precipitation are an important feature of mass exchanges between lake and atmosphere whether they be considered as being beneficial or detrimental to the biological system. A constituent of precipitation which can be considered as being of the former type is phosphorus which is an important limiting nutrient in lakes. The roles of winter cover in the phosphorus loading of lakes are used as an illustrative example here.

Whereas in summer, precipitation-derived phosphorus enters the lake body more or less directly during rainfall events, winter receipts are stored in its snow and ice cover to enter the water body rather abruptly during spring melt. In spring, as in summer, the lake receives such inputs from the land portion of its watershed as well as directly from the atmosphere. However, it can be argued that the direct inputs are disproportionately important in that they are not subject to the depletion which affects surface and subsur-

face flow from the land.

When a winter lake cover melts, the phosphorus within it enters the water body. Some of this phosphorus was derived from precipitation received during the preceding winter while the remainder was already present in the lake at freeze-up and was incorporated into the winter cover as the ice sheet evolved.

It is possible to calculate spring loading L, (mg m^{-2} of lake surface), in this case of total dissolved phosphorus, from each component of a winter cover or from the entire cover by means of the following expression (after Adams et al. 1978, English 1978).

$$L = \rho \cdot d \cdot [P] \quad (3)$$

Where: [P] = total phosphorus concentration (mg l^{-1})
 ρ = density of the component (layer) concerned
 (e.g., snow, white ice, black ice) g cm^{-3}
 d = thickness of the layer concerned, m

Using white ice (wi) on Coon Lake (see Table 1), at the peak of the 1975-76 season as an example:

$$L_{wi} = 0.877 \cdot 0.286 \cdot 0.064 = 15.716 \text{ mg m}^{-2} \quad (4)$$

Table 4 (after Adams et al., 1978, 214) indicates the distribution of phosphorus between the components of that particular winter cover in which there was rather more black ice (31 cm) than white ice (29 cm). It is interesting to note that the snowpack which was actually present on the lake at peak ice contained only 5% of the phosphorus while the white ice contained 62%. Thus a calculation of spring gains of phosphorus by the water body which assumed that the late winter lake snowcover contained the entire winter's receipts would be grossly in error.

Table 4

Mean Thickness, Density, Total Phosphorus Concentration and Phosphorus Supply from the Snow and Ice Cover of the Lake

Ice Type	Mean thickness (m)	Density (ρ) (g cm^{-3})	Mean Total Phosphorus [P] (mg l^{-1})	Loading (L) (mg m^{-2})	Loading as %
White ice	0.286	0.877	0.064	15.716	61.9
Black ice	0.309	0.900	0.030	8.430	33.2
Snow	0.102	0.210	0.059	1.253	4.9
			TOTAL	25.399	100.0

However, the 25.4 mg m^{-2} loading value in Table 4 represents all phosphorus released into the lake as a result of the melting of the winter cover. The net gain of phosphorus from winter precipitation can only be calculated by subtracting the entire black ice term and that proportion of the white ice phosphorus which was derived from beneath the ice during slushing. Precise calculation of this last term requires knowledge of the phosphorus concentration in snow which was incorporated into the white ice and of that in the water immediately below the ice during the slushing phase. The latter will presumably be a relatively high value (see below) as the water concerned is likely to be enriched in nutrients by the freeze-out process associated with black ice formation. It is conceivable that more than two-thirds of the white ice phosphorus is, in fact, re-cycled lake water phosphorus.

Table 4 is interesting in that it reflects the processes of winter cover evolution through the concentrations of phosphorus in each component, the influence of exsolution in the low black ice values, the concentrating effect of the slushing process, etc. The effect of the melting of each layer is different in that the black ice must have a relative diluting effect on the water body in comparison to the white ice. Detailed study of the evolution of the snow and ice cover of a lake and of the layer of water immediately below it should shed light on cycling of nutrients and other water constituents. For example, how significant are gains to the water body from exsolution during the winter? And, is the slushing process a one way street for nutrients or pollutants or is there a return flow of enriched slush water during white ice formation? Aspects of such topics as these are considered below.

The magnitude and proportions of the inputs presented in Table 4 will naturally vary greatly from case to case. Wolfe (1979, 1980) working on Lake St. George in 1977-78, reported concentrations of phosphorus of 0.013, 0.008 and 0.014 (mg l^{-1}) for white ice, slush ice and snow, respectively. Such variations reflect both differences in winter cover-forming processes, for example, the effectiveness of freeze-out and differences in lake chemistry. In a low snowfall situation, as in the case of the 1979-80 winter in southern Ontario, the snow-derived component will be negligible and the black ice ("recycled") component will be relatively and possibly absolutely greater. Also, in terms of the overall Spring loading of a lake, the relative significance of the lake cover-derived inputs will, of course, depend on the size of the lake surface in comparison to the size of its land catchment area (the A_c/A ratio of Table 1, see Schindler, 1971). Furthermore, the impact of the inputs on the lake as a whole will depend on the extent to which they are mixed into the water body. In some cases (see, for example, Wolfe 1979 and 1980), one effect of the ice cover is to preserve the thermal stratification of the winter so that spring melt and its constituents, particularly the land snowmelt which runs off relatively early, effectively passes out of the lake having only been associated with its surface layers.

Aspects of the detail of the interaction of the unfrozen water column and the atmosphere including the lake snowpack, through the white ice-forming process were considered by Orr (1980) and Pearson (1980) in connection with a study of major cations (calcium, magnesium, sodium and potassium) which form micro nutrients in the lake system. These are better indicators of ice forming processes as, in addition to existing in early detectable quantities, they, unlike phosphorus, exist in a single chemical form. This work brings out some interesting aspects of the white ice and black ice forming processes, especially with regard to the detail of sub-ice gradients and the implications of rapid and slow freezing situations on the freeze-out process.

It should be borne in mind that the processes of winter loading, (which is simply a facet of the mass balance of a lake) have wide implications for the understanding of the lake-atmosphere interactions. The currently topical subject of acid precipitation is an excellent example of this.

Winter Cover and Energy Balance

The initiation of a winter cover marks a drastic change in the energy balance of a lake in terms of the magnitude and nature of the components of the balance and in terms of the exchange processes involved. As was indicated above, mass exchanges between water and atmosphere are dramatically reduced and so therefore are energy fluxes associated with them and turbulent exchanges of sensible heat. The radiation balance at the lake surface and the radiative flux between the lake body and the atmosphere are also greatly altered. Once again, the nature of the cover, rather than the simple existence of a 'lid', becomes particularly important.

A layer of black ice is effectively transparent to radiation as a result of its relative purity and the vertical alignment of its crystals (Adams 1976a, Pounder 1965). Thus a lake covered with black ice may experience appreciable warming after freeze-up as a result of the efficient penetration of radiation into it in conjunction with the reduction in turbulent energy losses resulting from the presence of the ice sheet. Such lakes as, for example, in Antarctica or the High Arctic or in southern Ontario in 1980, may

also experience early warming in the spring. Black ice appears black because it reflects little radiation whereas snow appears white because it has a high albedo and is relatively opaque to radiation. Thus the development of a snowcover on a lake reduces inputs of radiation markedly but, because of its low thermal conductivity, it also reduces energy losses from the lake during the winter. White ice can be viewed as lying somewhere between the other two components (although closer to black ice) in its energy balance roles. In comparison to black ice, it has a high albedo but a somewhat lower thermal conductivity.

Thus, in spring, for example, as the normal sequence of melting is snow→white ice→black ice, areas with thick snow and white ice covers melt last, the pattern of snow and white ice accumulation may offset the common tendency for a shore moat to develop around a lake at the beginning of the melt. At such times, the snow and white ice, by lowering net radiation at the surface and by limiting the transfer of energy down the atmosphere→water temperature gradient, slow the warming of the lake. This is in contrast to the role of the same layers during cold phases of the winter when their principal effect is to limit heat losses along the water→atmosphere temperature gradient.

It is interesting, as an illustration of roles of the cover components in the energy regime of a lake, to consider the 'annual heat budget' of an ice-covered water body. This is a useful exercise in that it nicely brings out the important distinction between externally-derived components of the winter cover and internally-derived components. This distinction was, of course, of considerable significance in previous sections of this article.

The annual heat balance of a lake is its response to gains and losses of energy from all sources (see Scott and Ragotskie 1961). It can be considered as representing the total amount of heat necessary to warm up a lake from its lowest annual temperature to its highest annual temperature. In this context, the ice and snow cover of a lake is important not only in modifying lake-atmosphere energy exchanges but also as a store of latent heat. In spring, heat is required not only to raise the temperature of the snow and ice to the melting point and to warm up the unfrozen water but also to effect the change of state of the cover (Adams and Lasenby 1978). In some situations, the melting of snow and ice accounts for the major portion of winter heat income (see, for example, Schindler 1971).

Scott and Ragotskie (1961) express the heat budget of a snow-free, ice-covered lake as:

$$C_i + F_i = R_{si} + R_l + K_i + K_s + L + E + P \quad (5)$$

Where: C_i = change in sensible heat stored in ice below reference temperature of 0°C

F_i = change in amount of latent heat of freezing stored in the ice

R_{si} = incoming solar radiation absorbed by ice

R_l = net longwave radiation

K_i = sensible (turbulent or molecular) heat exchange through the water to the ice

K_s = sensible heat conducted from ice to snow

L = sensible (turbulent) heat exchange with the air

E = latent heat exchange with the air

P = sensible or latent heat added in rain, snow, etc. (above or below a reference 0°C so that latent heat of snow is negative in sign).

They express the heat budget of a lake snowpack as:

$$C_s + F_s = P + R_{ss} + R_l + K_s + L + E \quad (6)$$

Where: C_s = change in sensible heat stored in snow below a reference temperature of 0°C

F_s = change in amount of latent heat of freezing stored in the snow

R_{ss} = incoming solar radiation absorbed by snow

Only the terms P , F_i , and F_s are of interest to us here (see Adams and Lasenby (1978)).

During a winter, a lake may accumulate ice frozen in situ (our black ice) or formed of water frozen elsewhere and brought into the lake in the solid state. Examples of the latter include frazil ice, formed in open water upstream of the lake, ice bergs floating in a proglacial lake or snow lying on the lake or incorporated into its ice sheet as white ice. We are considering only the snow and white ice cases here but the fact that these are similar, in terms of their external derivation, to, for example, the ice bergs cited, is important. All ice which is in a lake in the spring, except any which escapes through the outflow, must melt in the lake. Energy is required, 79.72 cal g^{-1} (the latent heat of fusion of water) to effect the change of state in all cases.

Thus solid precipitation, P , arriving in the lake and melted there consumes energy which would otherwise be used for warming the lake body. In this sense, each snowfall (input of external ice) represents a loss of heat insofar as the heat budget of the lake is concerned. At a given time, the snowcover of the lake, F_s , represents a partial accumulation of these energy losses.

Obtaining the F_s term involves calculating the water equivalent of the lake snowpack (depth x density) and multiplying by the latent heat of fusion. Calculating the F_i term involves exactly the same procedure although ice density is less easily measured in the field (Adams and Lasenby 1978). However, for both conceptual and practical reasons, it is important to realise the limitations of these two terms as they have been described thus far and as they are commonly defined in the literature.

If there is white ice present, then the F_s term does not represent all inputs of snow (energy losses) since freeze-up as a greater or lesser proportion of snow receipts has been incorporated into the ice cover. If there is white ice present, it is also worthy of special attention if only because its density is likely to be different (generally lower and more variable, Adams, 1976c) from that of black ice so that accurate calculation of the F_i term requires knowledge of white ice density. In short, without a consideration of white ice, the F_s term is likely to be too small to represent externally-derived inputs and the F_i term too large to represent internally-derived inputs.

Examples of the calculation of cover component latent heat terms are given in Table 5.

Table 5

Sample Calculations of Latent Heat Terms

Volume x density x latent heat of fusion ($79.72 \text{ cal. g}^{-1}$)

that is: Water equivalent x latent heat of fusion

e.g., for Gillies Lake (see Table 6)

$$F_s = 20.81 \times 0.264 \times 79.72$$

$$\text{that is: } = 5.51 \times 79.72 = 439 \text{ cal. cm}^{-2}$$

$$F_{wi} = 36.47 \times 0.838 \times 79.72$$

$$\text{that is: } = 30.56 \times 79.72 = 2436 \text{ cal. cm}^{-2}$$

Values for F_i , the sum of white ice and black ice components are presented in Table 6. The F_i value for the entire ice sheet is given here as $4,225 \text{ cal.cm}^{-2}$ and $6,506 \text{ cal.cm}^{-2}$ for Gillies and Knob Lakes respectively. This value is, of course, considerably affected by the density used for the ice layer concerned. In the table, the snow term, F_s , is an underestimate in the sense that it does not include snow incorporated into the ice sheet. Adams and Lasenby (1978, 1027f) revise this value to include all snow receipts (Table 7). The new values, representing all external inputs of the winter, are more than double the original ones and the revised F_i values are approximately 70% of the original ones. The revised values more accurately reflect the role of the snow component of the lake cover as a consumer of heat.

Table 6

Calculation (depth x density x latent heat of fusion) of F_i and F_s (in cal.cm^{-2})
 F_i is partitioned into white ice and black ice

	\bar{h} (cm)	$\bar{\rho}$ (g cm^{-3})	Water Equiv. (g cm^{-2})	F_s	F_i
Gillies Lake					
Snow	20.81	0.264	5.51	439	
White Ice	36.47	0.838	30.56		2,436
Black Ice	24.94	0.900	22.45		1,789
Knob Lake					
Snow	41.37	0.315	13.07	1,041	
White Ice	42.11	0.886	37.31		2,974
Black Ice	49.22	0.900	44.30		3,532

Table 7

Calculation of Contribution (cal.cm^{-2}) of snow in white ice to winter budgets

	$\bar{\rho}_s$ (g cm^{-3})*	\bar{h}_{wi} (cm)	Water Equiv. (g cm^{-2})	Orig F_s	Revised F_s	Orig F_i^+	Revised F_i
Gillies Lake	0.264	36.47	9.63	767	439	1,260	2,965
Knob Lake	0.315	42.11	13.30	1,061	1,041	2,102	4,404

* If assumed that snow involved in slushing that produced white ice had same density as that present on lake at end of winter. By monitoring development of snow cover on a lake, it would be possible to use a measured value for each slushing event.

+ Slightly higher values for F_i here, as compared with Table 1, are a result of allocating densities between white ice and black ice (Adams 1976c:table 2).

Thus the various components of lake winter cover have distinctive roles in a lake's energy budget and an awareness of conditions and processes on and in that cover, throughout the winter, is necessary for a full appreciation of its energy regime. This is also true for a component of the energy balance of a lake, which is particularly important from a biological point of view, its light regime.

Winter Cover and Light Regime

The presence of a snow and/or ice cover greatly affects the light regime of the water body of a lake. Layers in the cover and interfaces between them, including the surface of the cover, control both the quantity and quality of light reaching the underlying water. The proportion of incident light which is transmitted, and the wavelengths involved are determined by the transparency of the layers and the degree of reflection and back-scattering at crystal and strata interfaces (Fig. 4).

Black ice has a low albedo and often is effectively transparent. It is not usually markedly stratified so that significant internal refraction is limited to zones containing bubbles or other inclusions or imperfections. The thickness of a black ice layer has little effect on light penetration. White ice and snow, however, have high albedos and because they are often layered and because crystals in them tend to be randomly oriented, they reflect, refract and absorb a high percentage of incident light. In this case, the thickness of the layers concerned is an important control of light regime. Snow is particularly important in limiting light penetration.

Roulet (1979) provides a valuable review of literature on this topic as well as some interesting measurements of percentage transmission and spectral distribution of light beneath varied snow and ice covers. The orders of magnitude of the effects of the three principal cover components discussed here are effectively conveyed by the statement by Maguire (1975, 1976), quoted by Roulet (1979, 13) that clear ice (black ice) allows transmission of more than 80% of total incident light as compared with less than 10% for white ice and less than 2% for snowcover. Roulet (1979, 71, Table 15) provides a summary of percentage transmission values from the literature. The dominance of snowcover, particularly of some types of snow, in the light regime of a lake in winter, is illustrated by Maguire's (1976) assertion that 10 cm of powder snow on a layer of clear ice may be sufficient to reduce light below photosynthetic levels. In the case of snow (particularly) and white ice, the type, number and orientation of crystals and the thickness and number of strata containing them are extremely important in determining their effects on light transmission. This makes generalizations about such effects more difficult than in the case of black ice. However, Roulet (1979, 72) makes the interesting observation that the relatively simple stratigraphy of the frequently "youthful" lake snowcover (see Adams and Prowse, 1978) simplifies relationships between light transmission and snow thickness on lakes. If this proves generally true, it may be that there is greater scope for establishing useful empirical relationships between light penetration and the simple measures of snowcover on lakes than on land where snow stratigraphy-radiation relationships can be extremely complex (see, for example, Giddings and LaChapelle 1961, Curl et al, 1972).

An example of Roulet's work on the spectral effects of lake winter cover is shown in Figure 5 (Roulet, 1979, 77, Fig. 14; see also Adams, W.A. 1978, and Adams, W.A. and Flavelle, 1977 for relevant studies). The diagram illustrates the marked effects of snow depth on the percentage transmission of light and also shows that there is a tendency for the upper and lower ends of the light spectrum to be least affected by the cover (here the variation in cover is mainly in the snow component). The greatest absorption is in the central, 500-600 nm range. The utilization of light by aquatic life is quite highly specific but the wavelengths 660-665 nm and around 430 nm are often cited as being most important for photosynthetic purposes.

Thus the winter cover of a lake can be viewed as a selective filter of light, controlling both the amount and type of radiation reaching the water body. The light which does penetrate through the cover is the principal control of productivity in the water. Variations in the thickness, spatial extent and temporal duration of one or all of the cover components considered here can have dramatic effects on a lake's winter light regime.

The role of the light regime in primary production by phytoplankton is developed by Roulet (1979) and is illustrated in Figure 5. This aspect of the influence of light in lakes is also treated by Pick (1977), Sagriff (1979) and others for lakes in east central Ontario. In Figure 5, the variation in carbon uptake below the ice (a measure of production) corresponds to the broad variation of light transmission which is in turn controlled

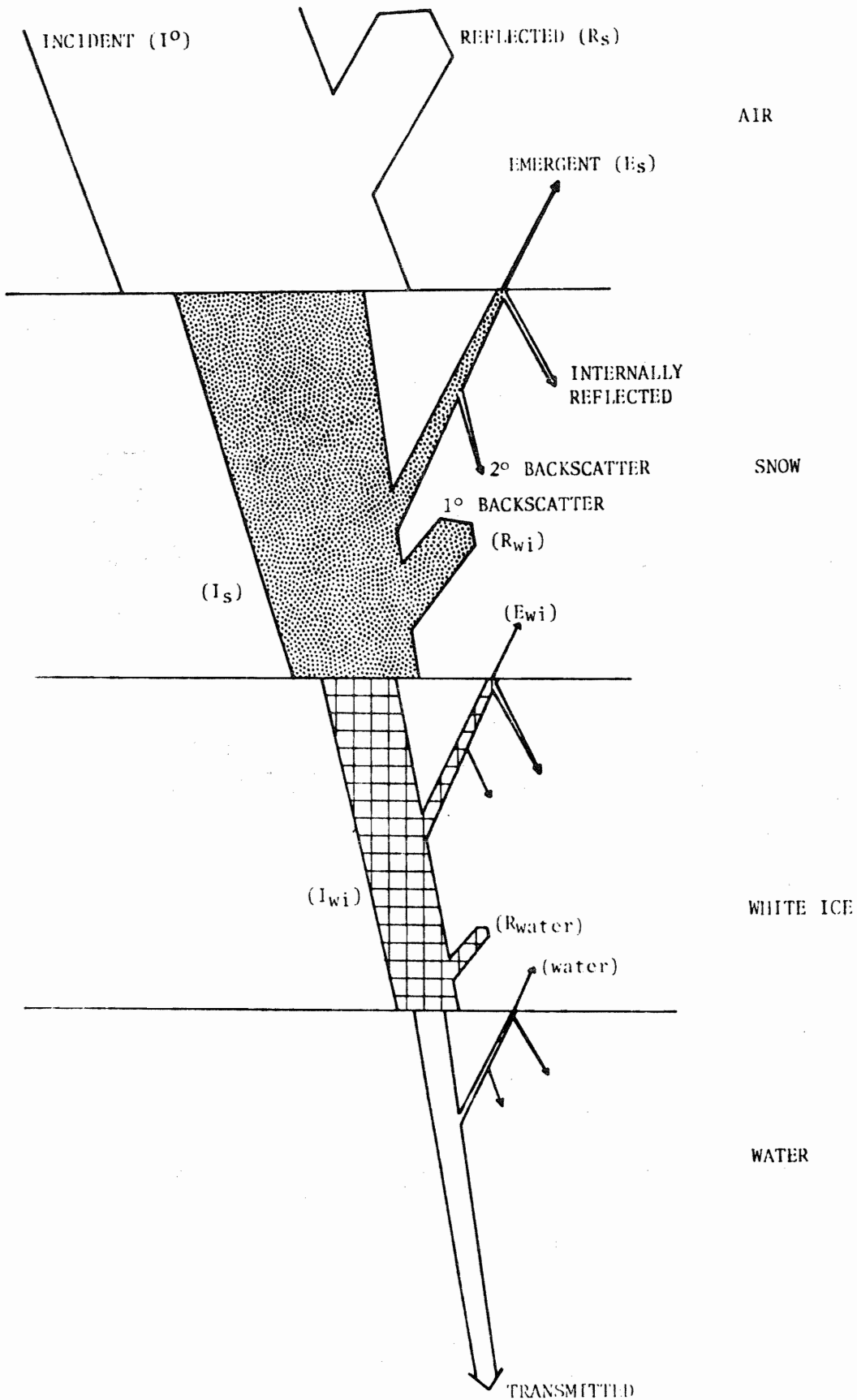


Figure 4. Roles of lake cover components on transmission of light. It is assumed that the black ice component is effectively transparent to light so that its behavior is analogous to that of water (from an article by Roulet and Wolfe in Roulet 1980).

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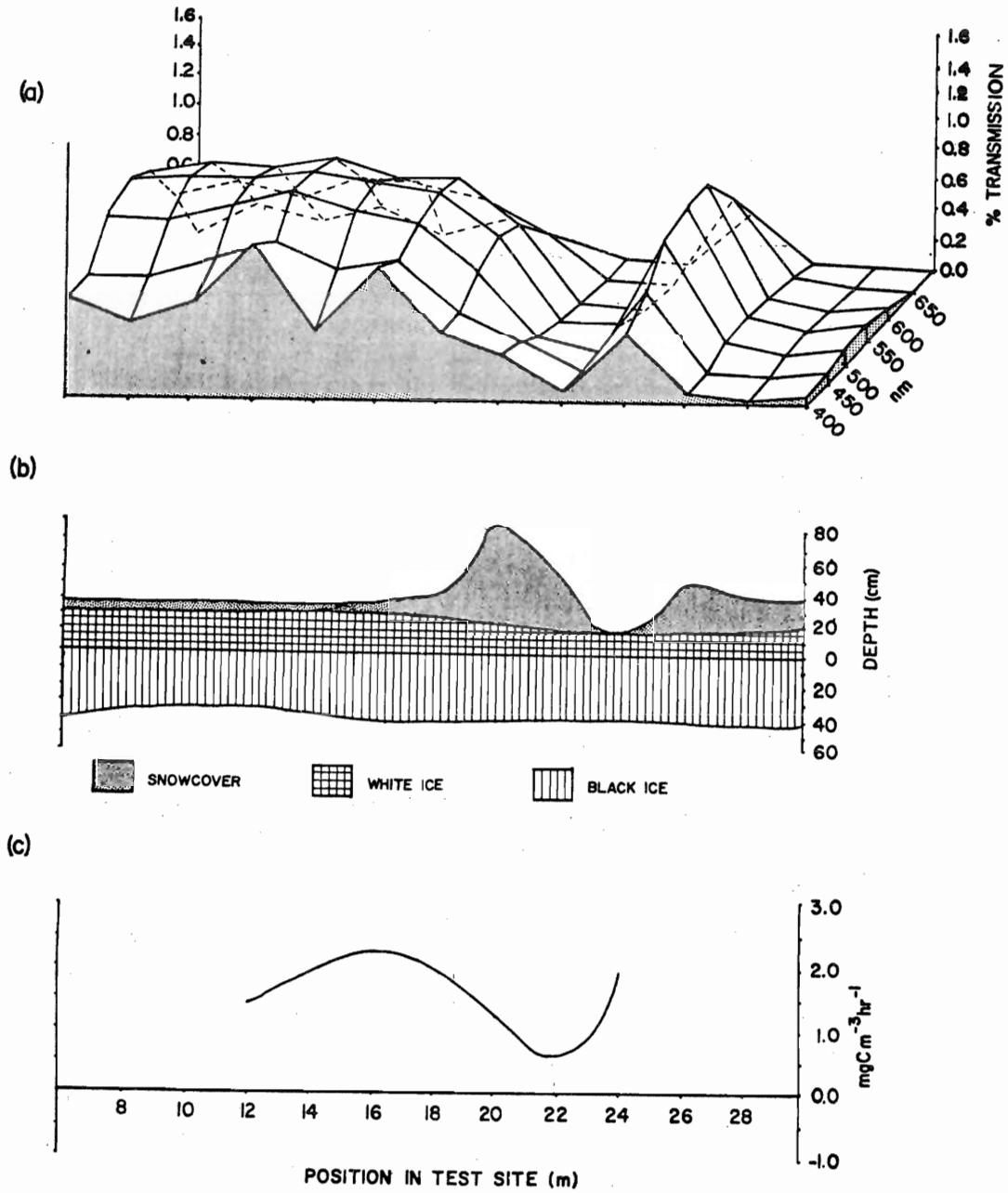


Figure 5. Variation in light transmission and carbon uptake along a strip of modified lake cover (from Roulet 1979).

by variations in the winter cover. The growth of algae and rates of photosynthesis have been shown to be directly related to light intensity, which, in winter, is controlled by the lake cover. Sagriff (1979) discusses the relationship between light intensity and the vertical migration of photoplankton under a developing winter cover.

Concluding Remarks

The purpose of this paper was to illustrate biological roles of the winter cover of lakes during research undertaken at Trent University, Ontario. It was not possible, in the time and space available, to encompass such diverse topics as the effects of ice-shove (see, for example, Adams and Mathewson, 1976), the roles of lake cover in seed dispersal and in animal movements during the winter or as a medium for algal life (see Adams 1976a). However, it is hoped that the examples provided are sufficient to indicate to Biologists that some detailed knowledge of lake cover would be a useful adjunct to their limnological studies and to students of snow and ice that there is more to the winter cover of a lake than "just frozen water".

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