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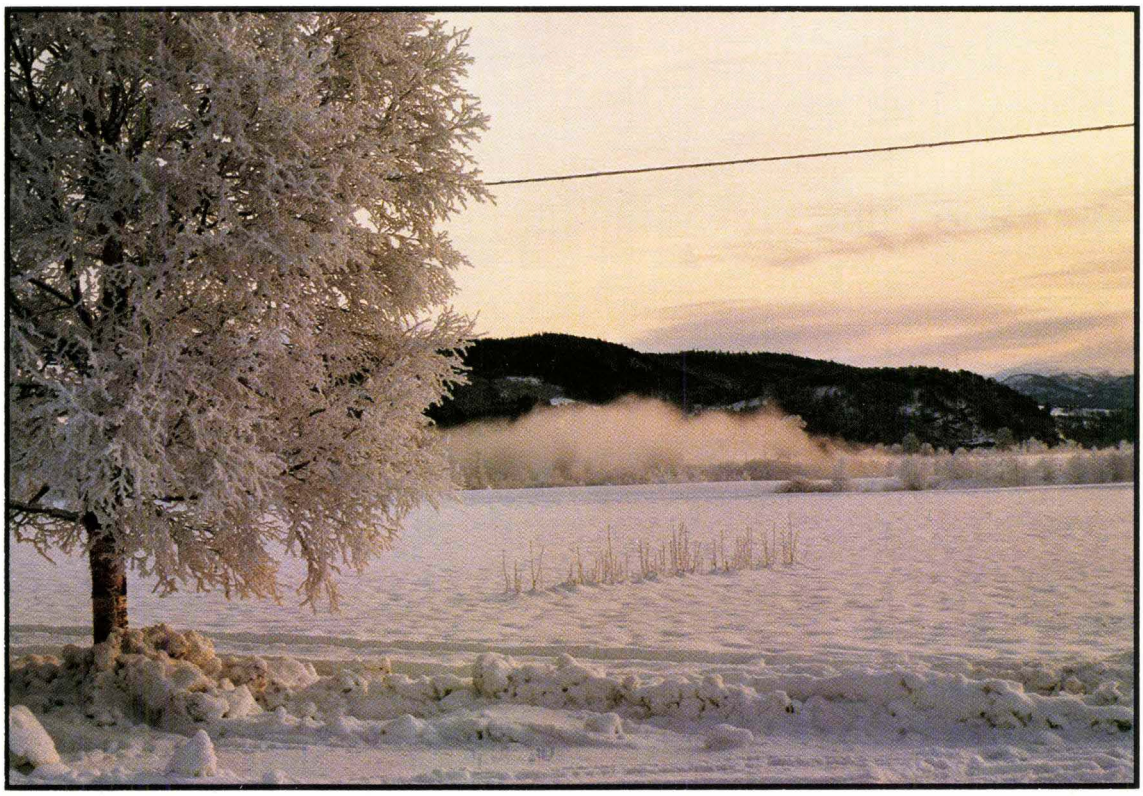
PUBLICATION

*Kåre Utaaker*

# EFFECTS OF WATERCOURSE REGULATION ON LOCAL CLIMATE



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A birch tree covered by hoarfrost,
and freezing fog above the Orkla River at Moan.
Photo: E. Skaar.

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ABSTRACT

Since the early 1960s, the need to learn more about the effects of watercourse regulation on local climatic conditions has given rise to numerous projects and studies. Despite this, few field studies have been conducted in watercourse before and after regulation. Most of the studies have comprised expert reports on climatic impact or papers delivered at seminars, etc.

In this report, the main emphasis is on results based on measurements and observations made over a number of years, both before and after regulation.

In Norway, most of the environmental impact of hydropower development occurs in valleys. Locale climate conditions in valleys exhibit special characteristics which are determined by water surface area, water level, water temperature and ice conditions in the lakes and rivers in the area. Watercourse regulation can therefore lead to significant changes in air temperature, humidity, freezing fog conditions and rime deposits along the watercourse.



SUBJECT TERMS

watercourse
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## *PREFACE*

Since the early 1960s, the need to learn more about the effects of watercourse regulation on local climate has given rise to numerous projects and studies on watercourses in Norway. Despite this fact, few field investigations have been conducted on watercourses before and after regulation. Most of the studies have comprised expert reports on climatic assessment done in connection with requests for concession for hydropower development, and with legal claims, or papers given at conferences and seminars. Frequently, the papers are based on local weather observations only after regulation, and/or on data from the regular network of weather stations operated by The Norwegian Meteorological Institute (DNMI), which may be more or less representative of the local climate under consideration. The results are also largely published in Norwegian in internal reports, and proceedings of conferences with rather limited distribution.

For some time there has been expressed a wish for a comprehensive survey of the results from the numerous studies. The present survey, which was commissioned by The Norwegian Water Resources and Energy Administration (Norges vassdrags- og energiverk, NVE), is based on a great number of publications. Some papers that do not contain result of special interest, are not listed in References. Interesting publications on the topic unknown to the author may also exist.

The main emphasis is put on results based on measurements and observations made over a number of years, both before and after regulation. However, also some studies after regulation, and some field investigations not directly related to watercourse regulations have added valuable documentation. The geographical positions of the research areas in Norway referred to in this survey are indicated on the map in Figure 1.

Some of the assessments and results given here are mainly based on principles of environmental physics, and general knowledge of micro- and local scale climates. This knowledge is also essential

for the understanding and interpretation of the results arrived at in field investigations. Simple explanations of some physical processes operating in the earth's surface layer, and of some small scale meteorological phenomena connected with watercourses are therefore given in Chapters 2, and 6.2 (Appendix).

In closing, I wish to thank Frank Cleveland for help in the preparation of the diagrams, and those who kindly gave permission to reproduce photograhs. Thanks are also due to my wife, Kirsten, for proof reading, and especially to Arve M. Tvede for valuable suggestions and careful proof reading.

Bergen, February 1995

Kåre Utaaker

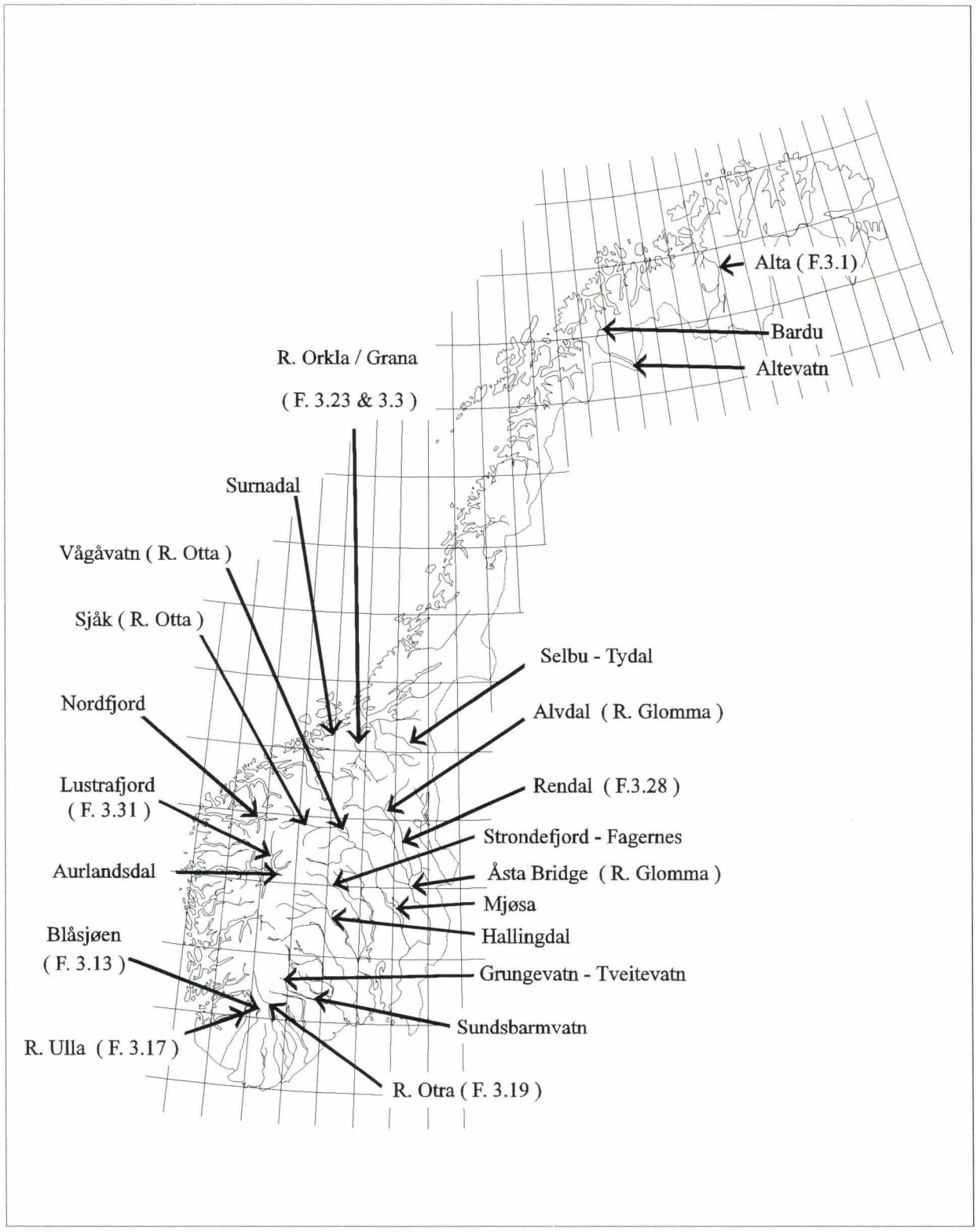


*Figure 1. Map indicating the geographical positions of research areas in Norway referred to in this survey. For some studies a detailed map is presented at relevant place in the text (e.g. F.3.28 = Figure 3.28, a map of Rendal).*

# *CONTENT*

# 1. SUMMARY AND CONCLUSIONS

In Norway most disruptions of nature caused by hydropower development take place in valleys. The local climate of a valley exhibits special characteristics which are partly determined by water surface area, water level, water temperature, and ice conditions in lakes and rivers in the area. Watercourse regulation can therefore lead to significant changes in ambient air temperature, humidity, fog conditions, and frost deposits along the watercourse.

***Damming of rivers and lakes*** changes the topography, and higher groundwater level in flat areas just above the highest regulated water level (HRW) may make the land marshy. When a reservoir is filling above natural level prior to regulation, area, volume, and heat capacity of the waterbody increase. Thus, diurnal and sesonal temperature changes, which are generally small in waterbodies, are dampened. Therefore, the effect of a reservoir on ambient climate varies with water level.

***In spring*** (after ice break-up) and ***early summer*** the rise in the surface temperature of a reservoir lags behind that of the area prior to regulation, and especially on warm days the water temperature is usually markedly lower than ambient air temperature. By light breeze areas exposed to onshore winds may experience a considerable lowering in daytime air temperature, and a rise in the temperature on cool nights due to the regulation (cf. Figures 3.4, 3.5, and Table 3.1).

***In autumn*** the cooling of a filled reservoir proceeds more slowly, and the ice freeze-up occurs at a later date than during similar preregulation conditions. Especially, in calm, clear weather this leads to higher ambient nighttime temperature, and the first frost is delayed in the new shore zone, whereas the daytime temperature is usually affected marginally (Figure 3.6 and Table 3.1). Delayed freeze-up also prolongs the season of more or less vigorous

evaporation from the waterbody. This may increase the incidence of fog (frost smoke) and hoar frost in the area.

***When the reservoir is covered with ice and snow***, possible changes in ambient climate are related to changes in topograhy and surface type. Flow and accumulation of cold air in the reservoir area will result in markedly lower temperatures in the new shore zone than before the construction of the dam, especially in calm, clear, or lightly clouded weather (Figures 3.7, 3.12, and Table 3.2). Since the area of the reservoir is usually much wider than the valley floor (the lake) before the damming, the cold air flow (sea) above the ice is more shallow than corresponding flow in this part of the valley before regulation. Moreover, a dam is frequently constructed at an constriction in the valley, which impeded the the flow of cold air more or less effectively. Therefore, the effect of a reservoir on ambient air temperature will always be restricted to a layer above HRW, which remains more shallow than the hight of the dam (Figure 3.8 and Table 3.2).

A water surface, and even more so an ice surface, is more smooth than most land surfaces. By a large increase in the area of a lake, and particularly by large artificial lakes, a considerable increase in the speed of onshore winds must be expected. The most striking changes occur where previously forested areas are inundated.

***Reduced waterflow in a river*** lowers the level and decreases the area of the water surface, and its temperature is also affected. The river and its shallow lakes are cooled and freeze up more rapidly in autumn. In spring ice break-up tends to occur later, especially in small lakes, weir basins, and slow flowing parts of the river. However, after the thaw of the ice, the temperature of the water rises more rapidly, and during summer it varies more in step with the air temperature, and tends to stay higher than prior to regulation.

***Lower water*** *level* may lower the groundwater level in the land along the river. This may affect the micro-climate of lowlying areas,

where water available to plants was, at least partly, supplied from the groundwater. When this supply is cut off, or reduced, desication of the soil during dry spells may cause higher daytime and lower nighttime temperature in the near-surface layer than before regulation (Figure 3.16). When marshy areas are drained the temperature changes can be fairly large. Lowering of the groundwater level on arable land may in some cases also lead to crop failure.

***The changes in the water temperature*** in spring affect the air temperature positively close to somewhat larger surfaces as e.g. weir basins. The effect is most pronouced on cool nights - particularly when the air temperture drops below 0°C. ***Reduction of the area of water*** has an opposite effect, and in some cases it can increase the incidence of frost on adjacent land. Daytime temperatures are rarely affected significantly by these changes in water area and water temperature.

In autumn rapid fall in water temperature, as well as reduced area may result in increased risk of frost near the river, whereas the daytime temperatures are hardly affected.

***Delayed ice thaw in spring*** can result in a marked lowering of the air temperature - mainly the minimum temperature - on adjacent land (Table 3.3 and Figure 3.20).

***Reduced discharge in a waterfall*** decreases or cuts off the spray precipitation characteristic of the climate in an area below the waterfall. The local air flow created by the falling water is also diminished or completely extinct. On warm summer days the daytime air temperature will be higher in most of the former spray zone, and on calm clear nights in autumn the absence of the "waterfall wind" may increase the risk of frost. However, the affected area is small at most waterfalls in regulated watercourses in Norway.

***Increased waterflow in a river*** usually rises the groundwater level in adjacent soils, and increases heat capacity and speed of the flow. Especially in winter, the flow downstream of power stations

increases considerably. Usually the discharge from a station is relatively warm in late autumn and winter and cold in summer.

In winter this results in no or less icing for some distance downstream of the station. In summer the water temperature stays markedly lower here than before regulation, especially on warm days, and where the remainig waterflow upstream of the power station is small compared to the outlet from the station.

In summer along rivers downstream of Norwegian power stations local climates are marginally influenced by these changes. On warm days cooled river air flowing onto the banks is mixed vigorously with warm, unstable air, and the effect of the cold water is rapidly diluted.

***Change in ice conditions due to regulation*** is the main cause of changes in local climates of Norwegian valleys. In calm, cold weather an open river downstream of a power station causes a marked rise in ambient air temperature (cf. examples from the Orkdal Valley p. 81). However, the most striking and visible effect is that vigorous evaporation from open water augments the humidity of the air, and hence increases the frequency and extent of frost smoke (Tables 3.4, 3.5, and 3.6), and the deposit of hoarfrost (Table 3.7). The increase in frost deposit is most marked in areas liable to invasion of frost smoke, and the amount usually decreases with distance from open water.

***Increased flow of freshwater into a fjord*** can strenghten, and enlarge the area of a stable surface layer of brackish water that frequently forms at the outlet of a river. In winter this can result in freezing over of a larger part of the fjord than in its natural state. By ice-covered fjord the shore zone experiences lower air temperatures than by open fjord under similar weather conditions (Figures 3.31 and 3.33. Effects of regulations are shown in Figure 3.34, and Tables 3.9 and 3.10.

***Detrimental effects of increased humidity of the air, frost smoke, and hoarfrost*** are common topics in legal claims related to development of hydropower.

***Frost smoke (fog) and heavy deposit of frost*** can most certainly be a nuisance to people who live, work or travel in areas exposed to these phenomena. In extreme cases road traffic may be impeded. However, it has not been possible to find scientific documentation for direct harmful effects of frost smoke or hoarfrost on human beeings, animals or plants. This applies also to possible damages due to heavy frost deposits onto outdoor wooden structures, painted surfaces and other materials.

Calculations evidence that increased humidity of the air and frost smoke do not aggravate ***moisture conditions in ventilated barns for housing of farm animals***, situated near an ice-free river. Moreover, experiments in the Orkdal Valley indicated that these changes in the local climate due to regulation had no measurable effects on ***moisture content of hay*** stored in an exposed barn.

# 2. SOME PHYSICAL PROPERTIES AND PROCESSES THAT AFFECTS LOCAL CLIMTE ALONG WATERCOURSES

## 2.1. Introduction

An axiom in modern physical climatology is that climates owe their individual characteristics to the nature of the exchange of heat, moisture, and momentum between the earth's surface and the atmosphere. This arises from the fact that it is mainly at earth's surface that short-wave radiant energy from the sun is absorbed and communicated to the soil as heat, and to the atmosphere as sensible heat, moisture (latent heat), and long-wave radiation. A small part of the long-wave radiation emitted from the surface is lost directly to space - more by clear skies than by dense cloud cover. Most of it is absorbed and converted to sensible heat in the atmosphere which in a complex process reemits long-wave radiation both downwards to the surface and upwards to space. Normally, the surface looses energy continuosly by long-wave radiation. When this loss is greater than the absorbed solar energy, the surface is cooled, and heat is "drawn" from the soil and air which both are subsequently cooled.

The energy exchange at the earth's surface which brings water vapour to the air and heats and cools the boundary layer, varies strongly with the type of surface and exposure. Local temperature and moisture conditions are therefore as a rule markedly influenced by the local surface types (forest, arable land, water, buildings, urban areas, etc.). The influence of a watercourse on ambient climate will depend on the area of lakes and rivers, discharge, waterfalls, water temperature, and ice conditions; and also strongly on the landscape. The local topography (mountains that obscure the sun, steepness and aspect of slopes, direction and width of valleys etc.) is as a rule the most important cause of local contrasts in the climate.

There is a saying that "micro-climate and local climate are fine weather climates". This relates the fact that small scale spatial contrasts, as e.g. in air temperature and air humidity, which are most fully developed in calm, clear weather are smoothend out in cloudy and windy weather. Therefore, the frequency of various weather events in the macroclimate is of major importance to the local variations in the climate of a region.

## *2.2. Some special features of valley meteorology*

In Norway, most of the disruptions of nature caused by hydropower development projects take place in valleys. In an attempt to map and explain the effects of the induced changes in the watercourse on ambient climate, it is frequently necessary to take into consideration special features of the valley climate.

Valleys produce their own local wind systems as a result of thermal differences. On calm days with clear or almost cloudless skies - especially in summer - the air above the sunny slopes and the floor of the valley is heated by the underlying surface to a temperature well above that of the centre of the valley. As a result a shallow, gentle, unstable upslope, and up-valley flow arises, usually a few hours after sunrise. Especially the up-valley flow, termed the ***valley wind***, increases gradually in thickness and speed and it normally lasts towards sunset. At higher levels, a closed circulation develops across the valley, together with a down-valley counter flow (Figure 2.1a). The exact nature of these wind systems depends on the orientation and the geometry of the valley. The best developed and most symetric system might be anticipated in a deep, straight valley with a north-south axis. In Norway, the wind speed 10 m above the valley floor rarely exceedes 4-5 m $s^{-1}$. The valley wind may also be combined with a large-scale breeze from the sea, the ***summer monsoon***, which develops e.g. when solar heating of the land results in a vast, weak low over the central part of Southern Norway.

At night the surfaces in the valley system cool by the emission of long-wave radiation. Thus the lower air layers are cooled, and since cold air is denser than warm air, the air near the surface slides gently downslope under the influence of gravity. The convergence of the ***drainage flows*** in the valley centre results in a weak lifting motion, and all of the downslope winds combine into a down-valley flow known as the ***mountain wind*** (Figure 2.1b). An up-valley counter flow is established aloft. The drainage flow often occurs as intermittent surges. Stable, cold air may be retarded or blocked by topographic constrictions and other obstacles in its path, and a more or less stable ***cold air lake*** is established. The coldest and densest air settles to the lowest levels producing an ***inversion*** where the temperature increases with height above the valley floor.

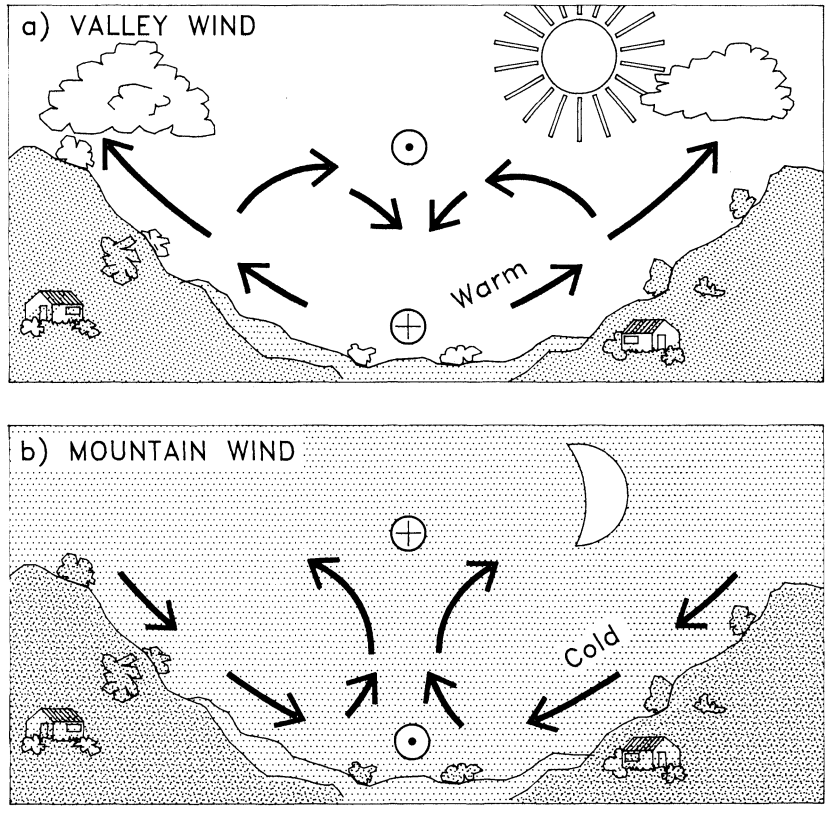


*Figure 2.1. a) The valley wind blows upslope and up-valley, ⊕ , near the surface and down-valley, ⊙ , aloft during daytime. The moutain wind blows downslope and down-valley, ⊙ , near the surface and up-valley, ⊕ , aloft.*

Under these conditions a more or less strong inversion will develop also in valleys with no marked impediment to the cold air flow, and the highest temperatures, the ***thermal belt***, are found on the valley sides at the level to which the cold flow builds up (Figure 2.2).

Cold air drainage, which may occur in all seasons, is normally strongest and most persistent in winter when snow covers the ground of the valley system. Especially in these situations an open water surface in the valley will influence the local circulation pattern markedly. The physical reason for this will be obvious from the discussion in the subsequent section.

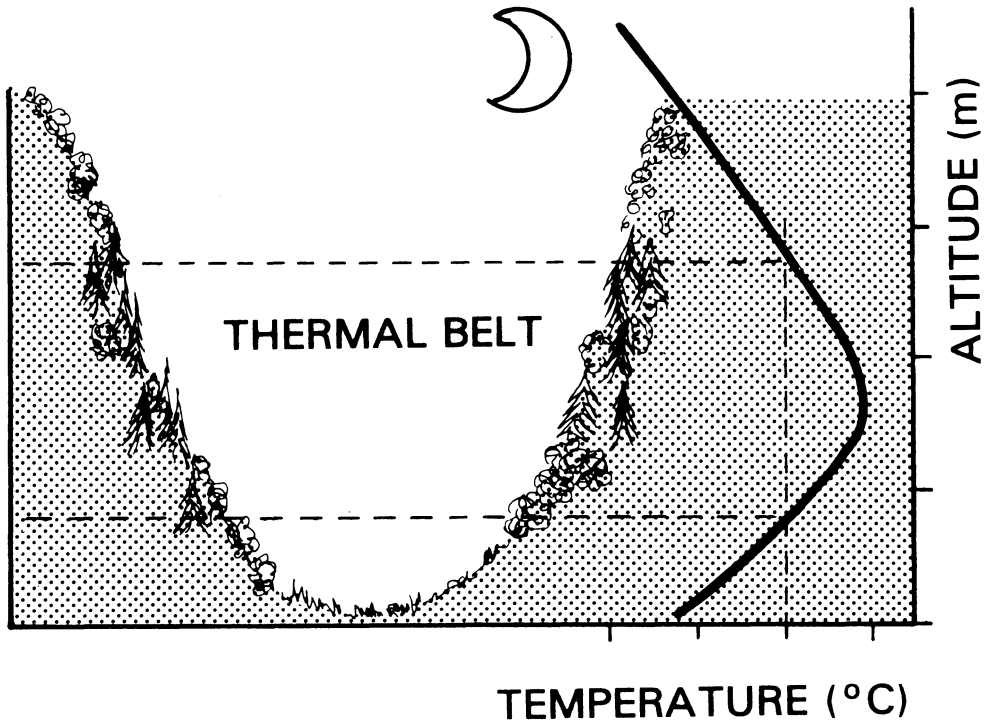


*Figure 2.2. On calm, clear nights cold air near the surface drains to the lowest lying portions of the landscape. The temperature curve to the right illustrates a typical vertical temperature distribution just before sunrise in a valley with a gently sloping floor along its axis.*

## *2.3. The air over water and ice surfaces*

Since regulations change the hydrology and the ice conditions of a watercourse, it is important to understand how the energy exchange at water and ice (snow) surfaces influences ambient climate.

The heat capacity of water is so large that it requires about three times as much heat to raise a unit volume of water through the same temperature interval as most soils. The solar radiation is asorbed within a large volume. Moreover, wind, and evaporative and radiative (long-wave) cooling of the surface create convection and mass transport by fluid motion that permit heat gains/losses to be spread throughout a large volume. Consequently, the thermal climate of a waterbody is remarkably conservative. This is evidenced in the diurnal (annual) heating/cooling cycle. Usually the diurnal (annual) ranges in surface temperature are far less at a water surface than at an adjacent land surface.

When warm air moves over a cold surface, it is cooled from below (Figure 2.3a). Sensible heat is transfered from the air to the surface. By weak wind a fairly stable, shallow layer will develop, and the air must travel a considerable distance before a thick layer is markedly affected. Therefore, the advective cooling of the downwind shores by rivers and small lakes will be restricted to a narrow zone. With increasing wind a thicker air layer will be affected, and the drop in the air temperature will be smaller. The surface temperature of ice is never above 0°C, and the cooling effect of a water surface with temperature above 0°C will therefore be less than that of an ice surface.

When cold air flows over a warm surface, it is heated from below (Figure 2.3b, and c). On watercourses this happens particularly over ice-free water during cold spells in winter, and in calm, clear nights in spring and autumn when air temperature above the land falls below that of the water. Since the warm air near the water surface is less dense than the cold air directly above, it

ascends. Vertical mixing results in a temperature rise in a fairly thick air layer, and consequently in higher air temperatures on the downwind shores (Figure 2.3b). When the air is relatively still, the flow of cold air from the surrounding land largely impedes the influx of warm air to the surface layer of the shore (Figure 2.3c).

Consequently, an open water surface dampens the range in ambient air temperature. Daytime (summer) temperatures are in general lower, and night (winter) temperatures higher in the shore regions than in more distant localities. The effect is largest on the downwind shores of large water bodies, and it decreases with decreasing water area, and with increasing distance from open water (cf. oceanic and continental climate).

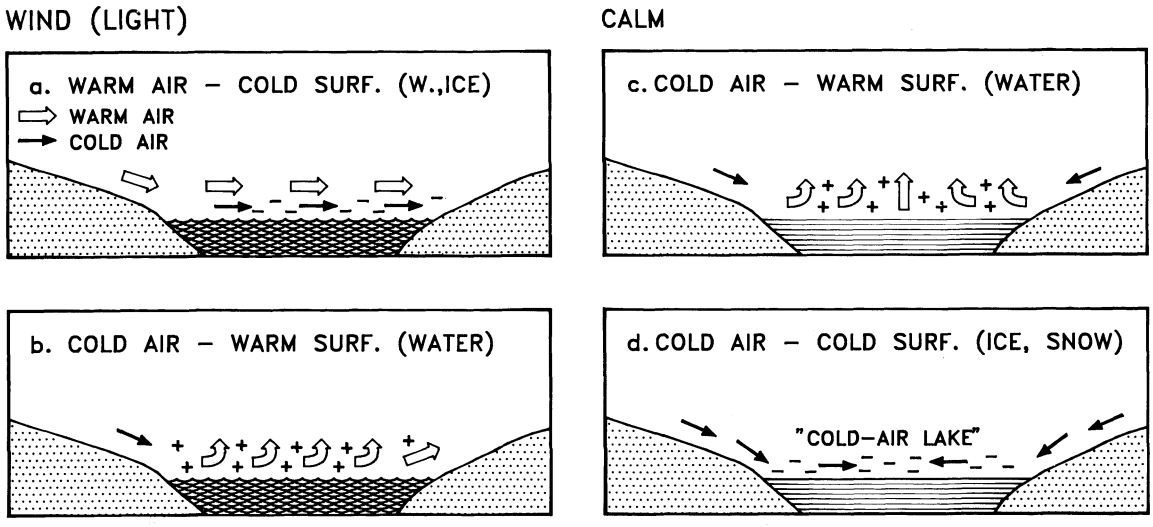


*Figure 2.3. Simple representation of the cooling, - , and the heating, + , of the air above water and ice surfaces.*

The thermal conductivity of ice is fairly high and that of snow is very low. Snow covering the ground acts as an insulating blanket. Hence, at low air temperatures the heat transfer from underlying water to the surface of bare ice will be markedly greater than to the surface of ice covered by snow. Very cold air that moves over bare ice may be slightly heated from below, whereas a similar flow

over snow-covered ice may be further cooled since the radiative loss from the snow surface markedly exceeds any heat conducted from the waterbody. During calm, clear nights extremely low temperatures can occur at the surface, and the build up of a strong inversion is a typical feature in these situations (Figure 2.3d).

Evaporation (transfer of water vapour and latent heat) from a water surface to the atmosphere will always take place when the air above is unsaturated, i.e. the relative humidity, RH , of the air is below 100% (cf. Figure 2.5). Table A2 in the Appendix gives estimated diurnal evaporation downstream of hydropower stations in Norwegian valleys under typical winter conditions. The calculations are based on a formula by Ryan et al. (1974). The table shows that the evaporation rate increases with increasing temperature difference between water and air, and with increasing wind speed. It is also higher by low than by high RH. However, at very low air temperatures the humidty of the air is of nearly no consequence.

In Figure 2.4 is depicted the energy loss by transfer of sensible heat, evaporation and long-wave radiation from a water surface with temperature 1°C by clear skies in winter. The evaporative loss, $Q_E$, is based on Table A2 (RH = 70%), the loss by sensible heat, $Q_H$ , on the Bowen ratio (Table A3), and the radiative loss, $L^*$ , is calculated by means of the formulas 6A and 7A in the Appendix. Note that the rate of increase in $Q_E$ and $L^*$ decreases slightly with increasing temperature difference between water and air, whereas the increase in $Q_H$ increases slightly. The total energy loss increases aproximately linearly with the temperature difference and with the wind speed. Clouds reduce the radiative loss markedly.

When cold air moves over warm water, evaporation will take place even if the air is saturated (RH = 100%). The air in contact with water is heated to the temperature of the surface, and accordingly RH drops below 100%, and water evaporates into the air. The warm, moist surface air is easily convected up into cooler air. If the mixture is super-saturated some of the water vapour condenses onto ***condensation nuclei*** (usually abundant even in clean-looking air),

and produces *fog* which visually appears like rising smoke (steam). When this phenomenon occurs over rivers, lakes, and fjords in winter it is called ***frostsmoke (steam fog, artic sea smoke)***.

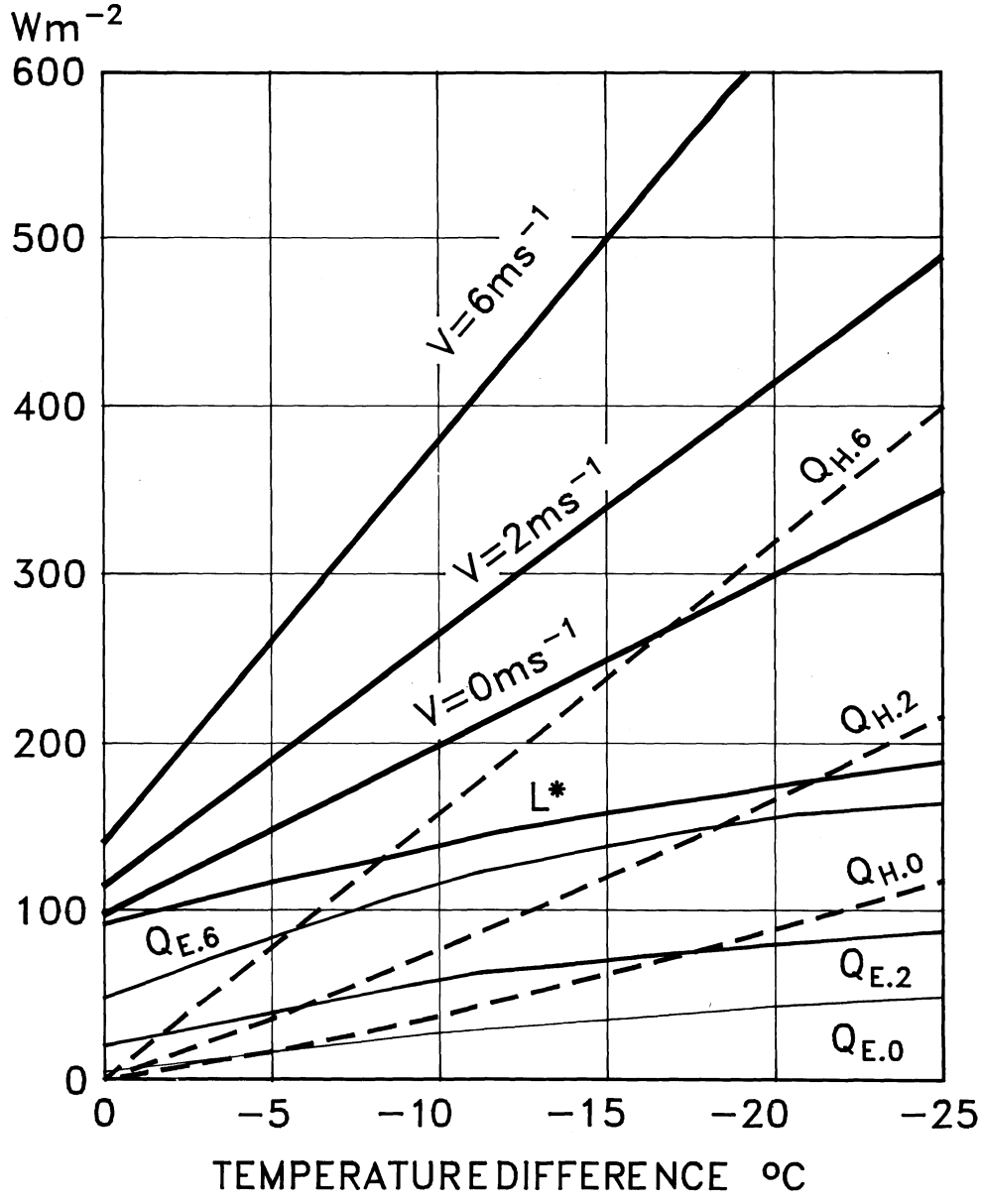


*Figure 2.4. Energy loss (W m$^{-2}$) to the atmosphere from a water body with surface temperature 1°C as function of wind, and temperature difference between air and water, by clear skies and no solar (short-wave) radiation. $L^*$ = loss by long-wave radition $Q_E$ = loss by evaporation (RH =70%), and $Q_H$ = loss by sensible heat. The subscript 0, 2, and 6 indicates wind speed (m s$^{-1}$).*

## 2.4. Frostsmoke. Formation and physical properties

In the ideal case, when deep, homogeneous, cold air overruns fresh, warm water of uniform surface temperature, a simple criterion for the onset of frostsmoke can be found by means of the vapour pressure versus temperature diagram in Figure 2.5 (Saunders, 1964).

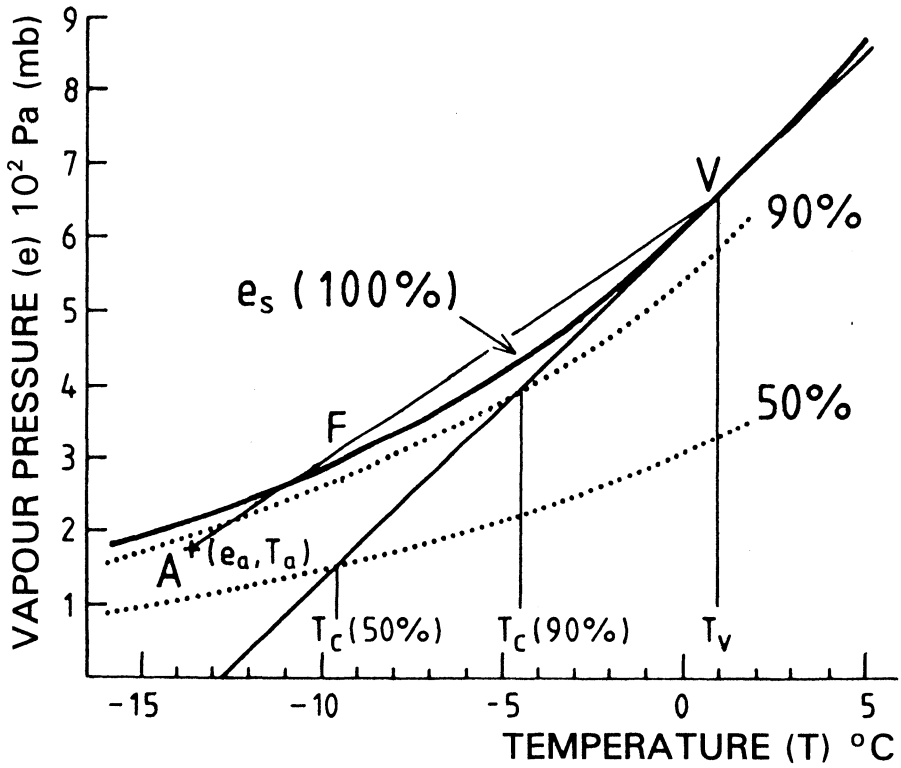


*Figure 2.5. Prerequisite for the onset of frostsmoke from fresh, water of uniform surface temperature, 1°C, when cold, homogeneous air overruns it. Explanations in the text.*

The cold, unsaturated air with temperature $T_a$ , and vapour pressure $e_a$ , advected from land is represented by point A, and the saturated air with temperature $T_v$ , and vapour pressure $e_s$ at the water surface is represented by point V. The mode of transfer and mixing of sensible heat and water vapour is nearly equal. The $(e,T)$ of the mixture of humid, warm surface air and the cold air above is therefore represented by the straight line between A and V. If this line intersects the curve of ***saturation-vapour-pressure***,

steaming may occur. Thus, points to the right of the intersection F on the line in Figure 5.2 represent super-saturated mixtures, and, provided the air contains ***condensation nuclei***, some of the water vapour condenses into tiny droplets. This implies that the tangent to the curve of saturation-vapour-pressure at the actual water temperature, $T_v$ , divides the $(e,T)$ plane in a significant manner (Figure 2.5). If the point $(e_a,T_a)$ lies above (to the left) of the tangent, steaming above the water may occur, if $(e_a,T_a)$ lies below (to the right), steaming cannot occur. The intersection between the tangent and the curve for observed relativ humidity of the cold air gives the threshold temperatur, $T_c$ , for just "no steaming". Figure 2.5 shows that for $T_v = 1°C$, $T_c = -4.5°$ when $RH = 90\%$, and $T_c = -9.6°$ when $RH = 50\%$. This indicates that the humidity of the cold air is of major importance when the temperature difference beteween water and air is small.

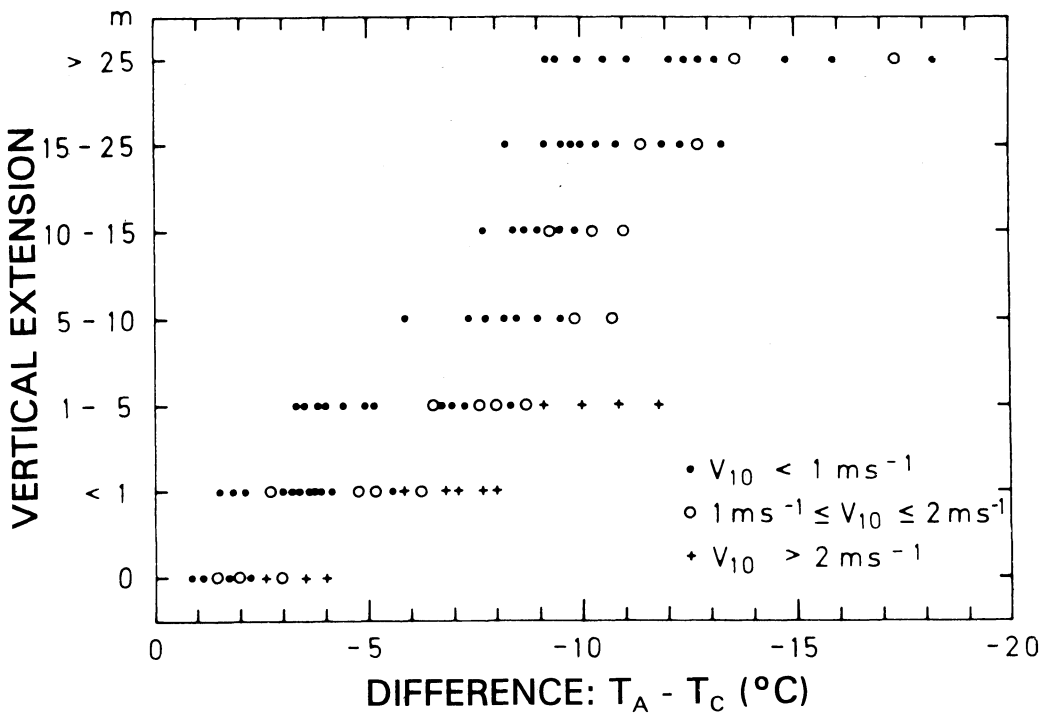


*Figure 2.6. The onset and the vertical extent of frostsmoke downstream hydropower plants in Norway in relation to the difference between air temperature, $T_a$ , and threshold temperature, $T_c$ , and wind speed, $V_{10}$ (Utaaker, 1979b).*

During cold spells in winter the circumstances by rivers and by small lakes downstream of power plants in Norwegian valleys are far from this ideal. Usually the cold air flow is an inversion layer of a depth of some 20-100 m. It is often erratic, and the leading edge of inflow from the surrounding ground is not easily defined. Nevertheless, the theoretical criterion depicted in Figure 2.5 can be a useful tool in an attempt to estimate expected frostsmoke from observations of $T_v$, $T_a$, $RH$, and wind speed.

Figure 2.6 which is based on observations from the rivers Glomma, Rena, and Otra gives the relation between $T_a$ - $T_c$ and the vertical extent of observed frostsmoke, grouped according to the speed of the cold air flow. Since frostsmoke often appears as feather-like plumes or small patches of steam at various heights, it may be rather difficult to appraise the vertical extent correctly. The height 0 in figure 2.6 means that the theoretical prerequisites for onset of frostsmoke were fulfilled, but that no visible "smoke" appeared. The observations indicate that the division between steaming and no steaming is at an air temperature some 1.5-2.0°C lower than the theoretical values. This can be explained by the fact that for the small "smoke plums" to be visible the air must contain a certain minimum amount of liquid water. To provide sufficient visual contrast in the poor daylight of northern winter a liquid water content of the order of 0.02 g $m^{-3}$ is a fair estimate, which would require a temperature, $T_a$, approximately 1.5°C lower than the threshold value, $T_c$.

The vertical extent and the density of the frostsmoke increases with increasing difference between $T_a$ and $T_c$. A difference of some -10°C is required to develop a layer of fog that reaches a hight of 15-20 m or more, and which spreads markedly beyond open water. It is, however, also dependent on the wind speed. Calm or very light winds, $V_{10} \approx 1$ m $s^{-1}$ or less, persisted during most of the observations depicted in Figure 2.6. With $T_v$ = 1°C, RH = 85%, and $V_{10}$ < 1 m s-1, which are typical values during cold spells, well developed frostsmoke that invades surrounding land is likely to occur when $T_a$ drops to -15°C or lower. Under these situations the

vertical and horisontal extent, and the density of the frostsmoke normally increase with increasing lenght of the cold air trajectory over the water (Figure 2.7).

At wind speeds about 2 m $s^{-1}$ or higher the entrainment of warmer and drier air from above, and/or from the sides is obviously so effective in Norwegian valleys that the frostsmoke normally dissipates rapidly above the water, and rarely spreads over land down-wind. Above wide water surfaces (lakes, fjords) widespread frostsmoke can develop also at higher wind speeds. This is illustrated in Figure 2.8. Under these circumstances a stratus cloud frequently develops at some distance down-wind of the leading edge (Figure 2.9). This is caused by gradual heating of the air in the surface layer, and hence diminishing evaporation (Table A2) which "lift" the condensation level in the cold air flow traversing the water.

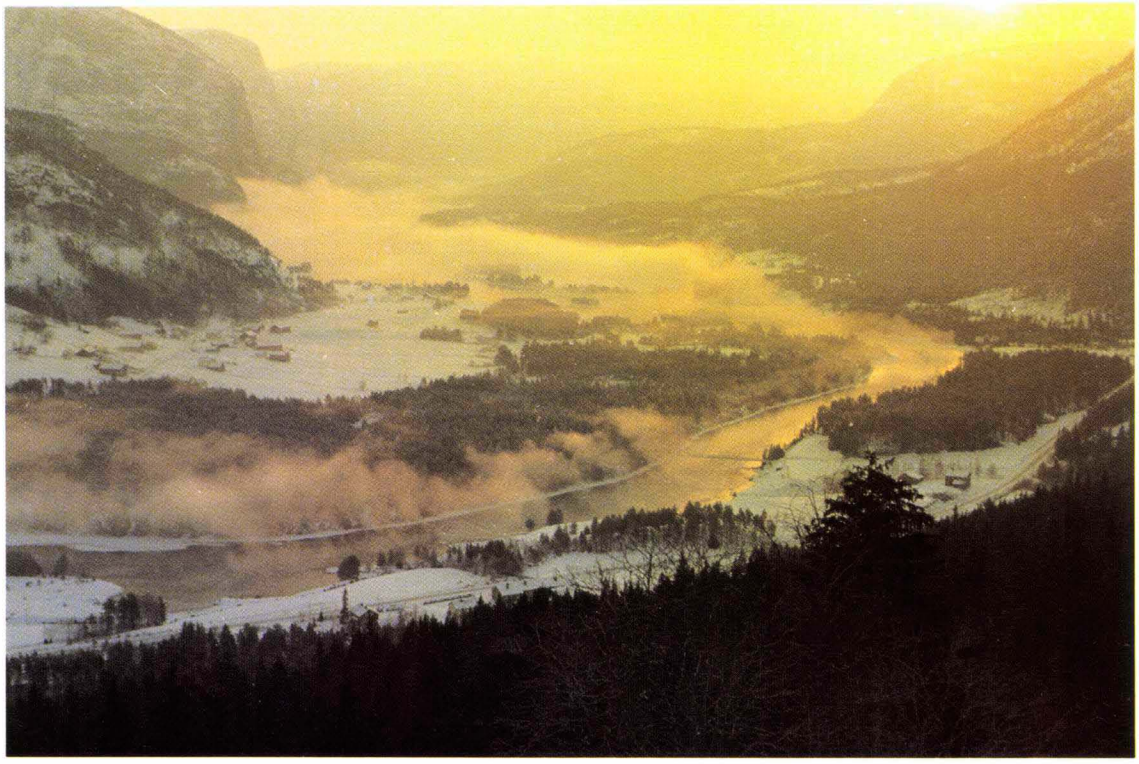

*Figure 2.7 Frostsmoke at Otra downstream of Brokke Power Plant. The temperature difference between water and air was 21°C, and the wind speed 0,6 m $s^{-1}$. The extent and density of the frost smoke increase markedly with the distance downstream.*

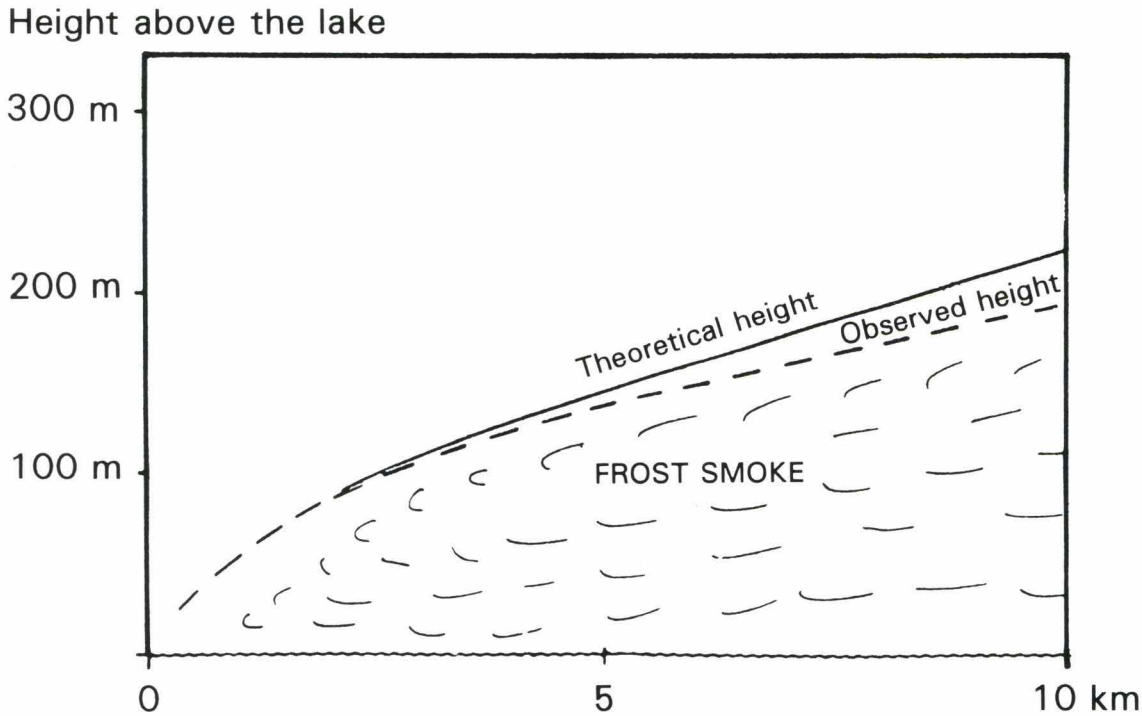


*Figure 2.8. Sketch of observed frostsmoke above Lake Strandefjord at Fagernes 29.11.87. $T_v$ - $T_a$ = 14°C, and wind: NW 4 m $s^{-1}$ (after Gotaas, 1988).*

*Figure 2.9. Stratus above Lake Storsjøen in Rendal 08.03.83. $T_a$ = -18°C, $T_v$ probably close to 4°C, and wind N circa 6 m $s^{-1}$.*

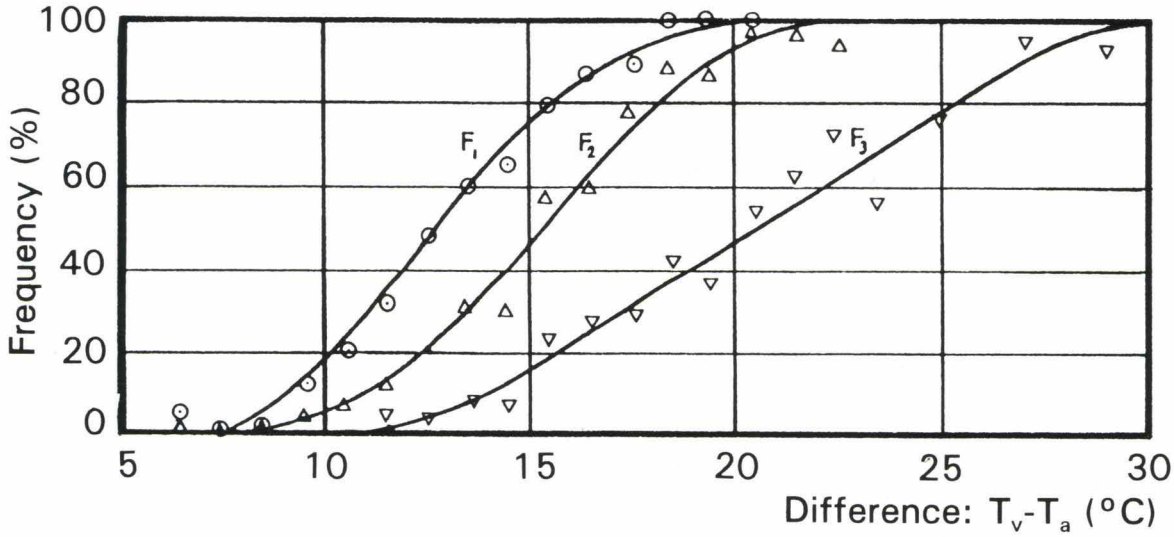


*Figure 2.10. The probability of frostsmoke in three classes: $F_1$, $F_2$, and $F_3$ as a function of the temperature difference between water and air. The classes are defined in the text (after Nordli, 1988).*

Based on observations throughout 14 winters at the outlet of Lake Vågåvatn, Nordli (1988) gives the probability of frostsmoke at an open water body as function of the temperature difference between the water surface and the advected air (Figure 2.10). The observed frostsmoke varied from barely visible steam patches just above the open water to dense fog that filled most of the valley volume. Possible radiation fog was not excluded. The probability curves in Figure 2.10 are given for the classes:

$F_1$ *All frostsmoke which can possibly be observed.*

$F_2$ *All well developed frostsmoke.*

$F_3$ *All frostsmoke that at least part of the time spreads beyond open water.*

Only $F_3$ can be a nuicance to people who live or travel in the area. The figure indicates that with a temperature difference of 15°C the probability of $F_3$ is about 15%, and with a difference of about 20°C

it increases to 50%. In this study the largest vertical extent of widespread frostsmoke (or radiation fog) was nearly always higher than 50 m, and in a few cases it reached about 300 m. During cold spells in winter wind speeds in this area rarely exceed 2 m $s^{-1}$. Therefore, Nordli did not study the effect of the wind on the onset and development of frostsmoke.

Provided all other conditions are the same both the vertical and horisontal extent of frostsmoke in Norwegian valleys will increase with an inreasing area of open water. At the outlet Lake Vågåvatn is usually ice-free during the first part of the winter. Later there is a lead about 200 m wide and 500 m long, and downstream some 4 km lenght of the circa 175 m wide River Otta is open. Average discharge in winter is 50-60 $m^3$ $s^{-1}$. The frostsmoke index developed by Nordli is likely to give realistic estimtes when applied in valleys with weakly developed mountain winds, and comparatively large areas of open water.

*Figure 2.11. Radiation fog, which is transformed to valley stratus, in Grindal by the Orkla 21.01.82 , 1130 CET. Air temperature about -6°C, and wind speed about 0.5 m $s^{-1}$ (photo E. Skaar) .*

The most extensive and dense local fog on open water in winter develops frequently by sudden change from mild, cloudy to clear, calm weather. Then radiative loss cools the ground rapidly, and if the air is relatively still, a shallow, moist surface layer is quickly cooled to super-saturation and ***radiation fog*** forms readily. If the air temperature falls to about -8°C or lower, evaporation from open water ($T_v$ = 1°C) becomes significant, and frostsmoke is added to the radiation fog.

Figure 2.11 shows a case of radiation fog in Grindal by the ice-covered River Orkla ***before*** regulation. By the sudden clearing of the sky in the night of 21.01.82 the air temperature dropped from about 2 to -8°C, and extensive, dense radiation fog developed, covering the entire valley bottom. When the picture was taken, the air temperature was -6°C, and the fog was dissipating. Hoarfrost on the ground and the forest was deposited from the fog (cf. Chapters 2.5, and 3.4.5), and the fog was transformed to ***valley stratus***.

The ***visibility*** in dense frostsmoke may be as low as 30 m, whereas in thin fog (mist) it may exceed 1 km. The content of ***liquid water (ice crystals)*** in frostsmoke may be estimated from observations of visibility. In fairly clean air most estimates are between 0.01 and 0.1 g $m^{-3}$, and in very dense frostsmoke about 0.2 g $m^{-3}$, whereas in strongly polluted air estimates are up to 0.4 g $m^{-3}$. In winter fog, in rather polluted air by the Angera River Baskirova and Krasikov (1958) measured 0.03-0.04 g $m^{-3}$, 0.05-0.11 g $m^{-3}$, and 0.08-0.37 g $m^{-3}$ in thin, moderate, and dense fog respectively. In the cleaner air above Kola Bay they found liquid water contents of 0.02-0.04 in moderate artic sea smoke, and 0.04-0,14 g $m^{-3}$ in dense sea smoke. At a pond of cooling water from a power plant in Arizona Currier et al. measured a maximum content of liquid water of 0.2 g $m^{-3}$ in very dense steam fog.

At higher air temperatures than -20°C the liquid water in frost-smoke is mainly ***super-cooled***, tiny droplets. Over Kola Bay Baski-rova and Krasikov (op. cit.) observed temperatures down to -22°C

before ice crystals were formed, whereas in the heavily polluted air over the Angera River they found ice crystals at -9 to -10°C, and here the condensate was mainly ice crystals at air temperatures below -20°C.

In both places the prevailing ***droplet size*** decreased, and the ***droplet specter*** narrowed with decreasing temperature, while the liquid water (ice) content increased. Currier et al. (op. cit.) found that the prevailing droplets had a diameter of 7.5-10 μm. Droplets of this size form rapidly by marked super-saturation, and they evaporate quickly when the relative humidity of the air drops somewhat below 100%. For example, a typical frostsmoke droplet, diameter 8 μm, will evaporate in 2 seconds if RH suddenly drops to 95%. This explains why frostsmoke dissipates rapidly by entrainment of relatively dry air from above and/or from the sides, and when ***hoarfrost*** is deposited onto the ground or onto trees and other objects within the fog.

## *2.5. Deposition of hoarfrost and rime*

Onto objects with lower temperature than 0°C water vapour is deposited (sublimated) as hoarfrost when air in contact with the cold surfaces becomes slightly super-saturated. Table 1A (Appendix) shows that at the same subfreezing temperature the saturation vapour pressure is smaller over an ice surface than over a water surface. This implies that hoarfrost can be deposited even from air with relative humidity, RH , below 100%. For example, if the surface and the air temperature both are -10°C, hoarfrost may form on the surface when RH is above 91%. Therefore, at subfreezing temperatures water vapour is "drawn" to the surface from saturated air over ice (snow, hoarfrost), and the relative humidity of the air drops. Tiny fog droplets in the surface layer eventually evaporate quickly and excess water vapour is continually deposited as hoarfrost on the surface.

Also when the watercourse is frozen up, hoarfrost is a common phenomenon in Norwegian valleys, espcially by sudden changes from mild, cloudy to calm, clear weather. An open water surface is an additional source of water vapour, and during cold, relatively calm spells hoarfrost is usually deposited in its vicinity even when no frost smoke occurs. The sublimation is largest in the down-wind "shore zone", and the amount of frost deposit usully decreases markedly with the distance from open water. Under such conditions an inversion prevails in the surface layer, and the deposition of frost from the moisture transferred from open water is limited to areas within this layer. When the air is relatively still (weak or no down-valley wind) the most marked deposits of frost appear frequently in a belt somewhat up the sides of the valley adjacent to the open waterbody. Warm, moist air ascends from the water and flows towards the slopes near the top of the cold air layer (cf. Figure 2.3c).

From extremely dense frostsmoke - e.g. at cooling ponds by very low air temperatures and moderate wind - the tiny, supercooled droplets may freeze on contact with objects in the flow, and accrete as a light, porous ***rime***. Downstream of Norwegian hydropower stations noticeable amounts of rime may occur solely on objects close to open water.

# 3. CHANGES IN LOCAL CLIMATE CAUSED BY REGULATION OF WATERCOURSES

## 3.1. Introduction

The development of hydropower changes the hydrography of a watercourse. By the construction of dams runoff of precipitation and meltwater from snow in the catchment area of the reservoirs is accumulated - especially during summer and autumn. Most of the stored water is used in the power stations in winter when the need for electric energy is usually large and the inflow to the reservoirs relatively small. The flow in waterways between intake and discharge areas for power stations is usully reduced strongly throughout the year. Downstream of power stations the flow is as a rule increased considerably in winter and decreased in summer. The discharge can vary markedly both between day and night, and between weekdays and holydays. Consequently, the pattern of the drainage of fresh water into the fjord is changed. In this context the increased discharge in winter after regulation is important. Flooding can also occur in regulated watercourses, but subsequent to regulation maximum flow is as a rule reduced considerably.

It is mainly the changes in area, level, temperature and ice conditions of the water surfaces of rivers and lakes - brought about by the regulation - that affects the local climate along a watercourse. Disrupture of local topography by construction of dams can also induce significant effects. Usually, the dam itself is the main agent, whereas construction roads, tips of rock excavated from tunnels, and quarries for material to the dam are of less importance. On fjords it is mainly changes in ice conditions that can affect the local climate.

Here examples will illustrate how disruption of nature caused by the development of hydropower can change local climate. The main emphasis is on some case studies where systematic measurements and

observations were conducted both ***prior*** and ***subsequent*** to regulation of Norwegian watercourses. However, our factual knowledge of the effects on local climate is also partly based on general knowledge of micro- and local climate, and on measurements and observations during a period ***after*** regulation only.

## *3.2. Reservoirs*

The damming of rivers and lakes changes the local topography, and when the reservoir is filled to highest regulated water level (HRW) the groundwater level in adjacent land can be raised, and flat areas just above HRW may turn marshy. Area, volum, and heat capacity of the waterbody increase with increased filling above natural water level before regulation. The effects of the waterbody on ambient climate will therefore vary with the filling of the reservoir and with the surface temperature, which partly will also depend on the filling. When the waterbody is covered by a sheet of ice the topographic effect predominates.

At an enormous reservoir (15.500 $km^2$) in The Sovjet Union, Venderov and Malik (1964) found that in extreme cases in summer ambient maximum air temperature is lowered about 3.5°C, and the minimum temperature is raised about 4.5°C. The frostfree season (mainly in fall) is prolonged by 5-15 days in the shore zone. The lake effect is largest in the downwind shore zone, but significant changes were recorded at distances up to 10 km from the reservoir.

At the ***Lokka Reservoir*** (417 $km^2$) in a flat landscape in Finland, Fransila and Järvi (1976) showed that the regulation had caused a rise in ambient minimum air temperature in fall and early winter (before the freezing of the lake). On average the rise was around 1.5, and 0.9°C at distances 100 and 400 m from the shore, and measurable changes were recorded up to 1 km downwind from the lake.

Some 50 m from the water of the ***Ljusnedal Reservoir*** (23 $km^2$) in Sweden, Rhode (1968) found an increase of 0.5-1.0°C in the minimum temperature on cold, calm nights in fall, and a similar decrease in the maximum temperature on warm days in spring and in summer.

In Norway the inundated area of hydropower reservoirs is mostly far less than in the example from Sweden.

***Lake Sundsbarmvatn*** in Telemark is dammed to HRW 612 m a.s.l., i.e. 20 m above natural water level (NW) prior to regulation. The lowest regulated water level (LWR) is 574 m a.s.l. The area of the lake which before regulation was some 5.9 $km^2$ is extended to some 8.8 $km^2$ at HRW. On a fairly steep slope at the small farm Sæli, The Norwegian Meteorological Institute (DNMI) operated 3 regular climatic stations (I:595, II:616, and III. 655 m a.s.l.) during 5 years before the filling of the reservoir was started in the summer of 1969 (Bjørbæk, 1970). Two of these stations (II and III) were in operation until the summer of 1974. Monthly summaries of the observations are published in the Yearbooks of DNMI (1964-74). Utaaker (1983) evaluated some of the original recordings, and the main results are given here.

Unfortunately, no systematic observations of the ice cover of the lake were carried through subsequent to regulation, when three of the winter were also exceptionally mild. It could, however, be ascertained that the reservoir - which is normally frozen up by mid-December - was covered entirely by ice at the end of January during all winters. In February a clear tendency to a lowering of the minimum temperature, $T_N$ , at station II compared to that at station III was found after regulation. Then the average water level (filling) was some 14 m below HRW (6 m above NW), and the mean drop was some 0.6°C in the temperature groups $T_N \leq -15°C$, and $-15°C < T_N \leq -10°C$ . In March with average water level at NW there was no tendency to changes caused by the dam, and in April with average water level 4 m below NW, there was a weak tendency to a relative rise in $T_N$ at station II by low temperatures. This indicates that the dam has not influenced the cold air buildup above

the frozen lake in the Sæli area when the water level is not above NW. This may be explained by the fact that local topographic constrictions in this part of the lake probably formed rather effective barriers to cold air drainage also before the dam was built. Therefore, observations of $T_N$ at the Sæli stations are hardly representative of the change in inversion buildup above the more open parts of Lake Sundsbarmvatn caused by the dam.

In the ice-free season there is a marked tendency to a relative rise in $T_N$ at station II in October and November (average filling 4 m below HRW) at air temperatures below 0°C only, with a significant mean difference of 0.45°C for this group in November. An influence of the dam on the maximum temperature, $T_X$ , is indicated during the summer months June-August only. A small, but statistically significant relative drop in $T_X$ has occurred at station II on warm days, with averages 0.2, 0.25, and 0.2°C for June (filling 11 m below HRW), July (filling 6 m below HRW), and August (filling 4 m below HRW) respectively.

This study indicates that the regulation of Lake Sundsbarmvatn has caused small changes in ambient air temperature. The influence on air temperature of an open or ice-covered waterbody decreses with increasing distance from the shore. Therefore, somewhat larger effects may be expected when the reservoir is filled closer to HRW than was the case for most the months during the postregulation period of this study. Moreover, a minor effect of the dam on air temperature at the reference station 45 m above HRW cannot be excluded.

In connection with the regulation of the ***Alta River*** some experts firmly stated that the construction of a some 120 m high hydroelectric dam in the narrow canyon at Savcu would induce large and harmfull changes in the local climate of Masi some 30 km upstream of the dam (Figure 3.1).

The argument was that the dam, which lifts HRW of Lake Virdnejavre 15 m (to the natural level of Lake Ladnatjavre), in cold air

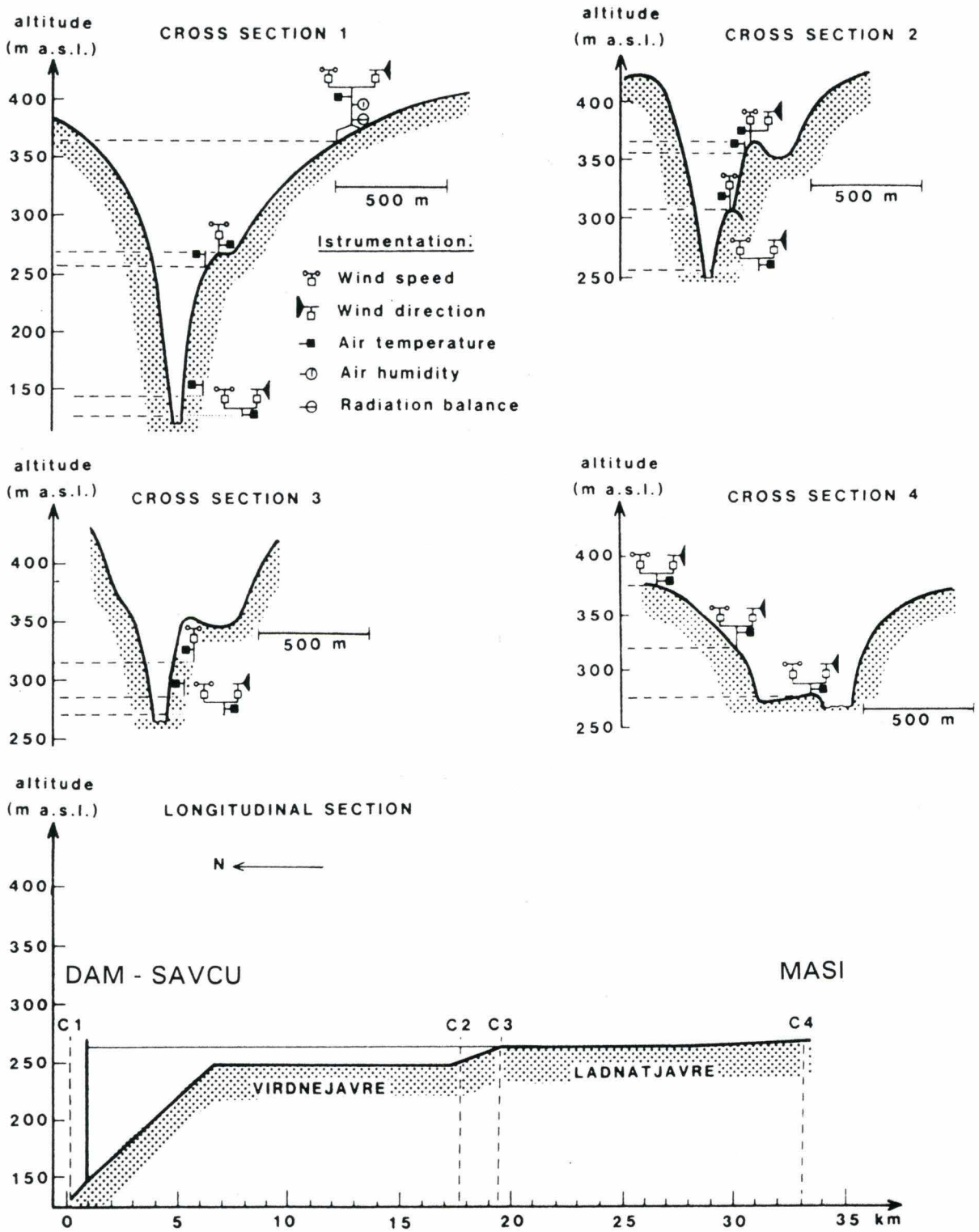


*Figure 3.1. Cross section in the longitudinal direction of the Alta Valley between Savcu (dam) and Masi, and some cross sections where also site and instrumentation of automatic, climatic stations are indicated (Gjessing and Hanssen-Bauer, 1988).*

drainage situations would impede the flow through the canyon between the two lakes (Figure 3.1, cross section 2 and 3). The result would be increased cold air stagnation above Lake Ladnatjavre and a marked lowering of the air temperature at Masi, particularly in spring.

To assess the effect of this change in hydrology and topography extensive field investigations were conducted both ***before*** and ***after*** the construction of the dam, which took place mainly in the summer and fall of 1986. But the reservoir was not filled until the spring of 1987. In Masi DNMI has three automatic weather stations in operation since 1981 (Figure 3.1, C4). In the project "Cold air production and drainage on the Finnmarksvidda Plateau" the Geophysical Institute, University of Bergen made detailed field measurements during the winters (October 15 through May 15) of 1984-88 (Gjessing and Hanssen-Bauer, 1988a,b, and Hanssen-Bauer, 1988b). The localization and instrumentation of automatic weather stations are shown schematically in Figure 3.1, C1, C2, and C3. In addition wind was recorded in Masi by means of a doppler sodar, air temperature was recoded from an airplane, and vertical temperature and wind profiles were measured by tethered balloon. Some additionel measurements were also conducted by Gotaas and Nordli (1990), who are engaged as experts on climate in connection with legal claims.

Hanssen-Bauer (1988b) presented results of measurements and model calculations. She conluded that the construction of the dam has not caused marked changes in wind speed or stagnation of cold air in the valley upstream of the dam. Model calculations for cold air drainage situations, with ice-coverd reservoir filled to HRW (i.e. the fall in the canyon between the lakes is eliminated) indicated a maximum lowering of some 0.3°C in the air temperature upstream of the canyon, and some 0.15°C downstream of the canyon. This insignificant effect disappeared when the reservoir was drained to natural water level prior to regulation, or lower. Evaluating recordings from the stations in Masi, Nordli and Gotaas (1993) found no significant effects of the regulation on the climate of Masi, a result which evidently validates the model calculations.

*Figure 3.2. Lake Granasjøen at Nerskogen.* ***Above:*** *The area before the construction of the dam.* ***Below:*** *The artificial lake is filling, summer 1981 (photos Å. Killingtveit).*

Next the effect of an artificial lake at Nerskogen (circa 63°N, 10°E) on ambient air temperature will be discussed in more details. ***Lake Granasjøen*** (Figure 3.2) covers an area of some 6.9 $km^2$ when it is filled to HRW (650 m a.s.l.). ***Before***, ***during***, and ***after*** the construction of the 45 m high dam, extensive meteorological field measurements were made in the area (Skaar, 1986). Map of the area with localization of climatic station is given in Figure 3.3.

### *3.2.1. The growing season at Lake Granasjøen*

On the average the growing season at Nerskogen (650-800 m a.s.l.) lasts from mid-May towards the end of September. The effect of Lake Granasjøen on ambient air temperature during this periode is evaluated by Skaar (1986).

The transformation from entirely snow-coverd to bare ground in the surroundings of the reservoir normally occurs in ***May***. Skaar found that the year to year change in this process was the main cause of recorded changes in ambient temperature from before to after regulation during this month. The interpretation of photos taken during April-May, 1979-85 produced no evidence of effects of the reservoir on snow melt in its surroundings.

Meltwater gradually fills the reservoir. As the water level rises, the sheet of ice breaks up, and the water surface is more or less covered by an ice floe until around May 20-25. When ambient air temperature is markedly higher than the surface temperature of the reservoir (about 0°C during ice melt), the downwind shores will experience a lowering of the temperature. By temperatures markedly below 0°C there is a rise in ambient temperature due to regulation. Provided all other conditions are the same, the temperature changes in a zone above HRW will increase with ascending water level. The reservoir has undoubtedly caused a lag in the seasonal rise in temperature in its immediate environment after the ground is free from snow in May. However, the collected data give no basis for quantifying the lag, which will also vary from year to year.

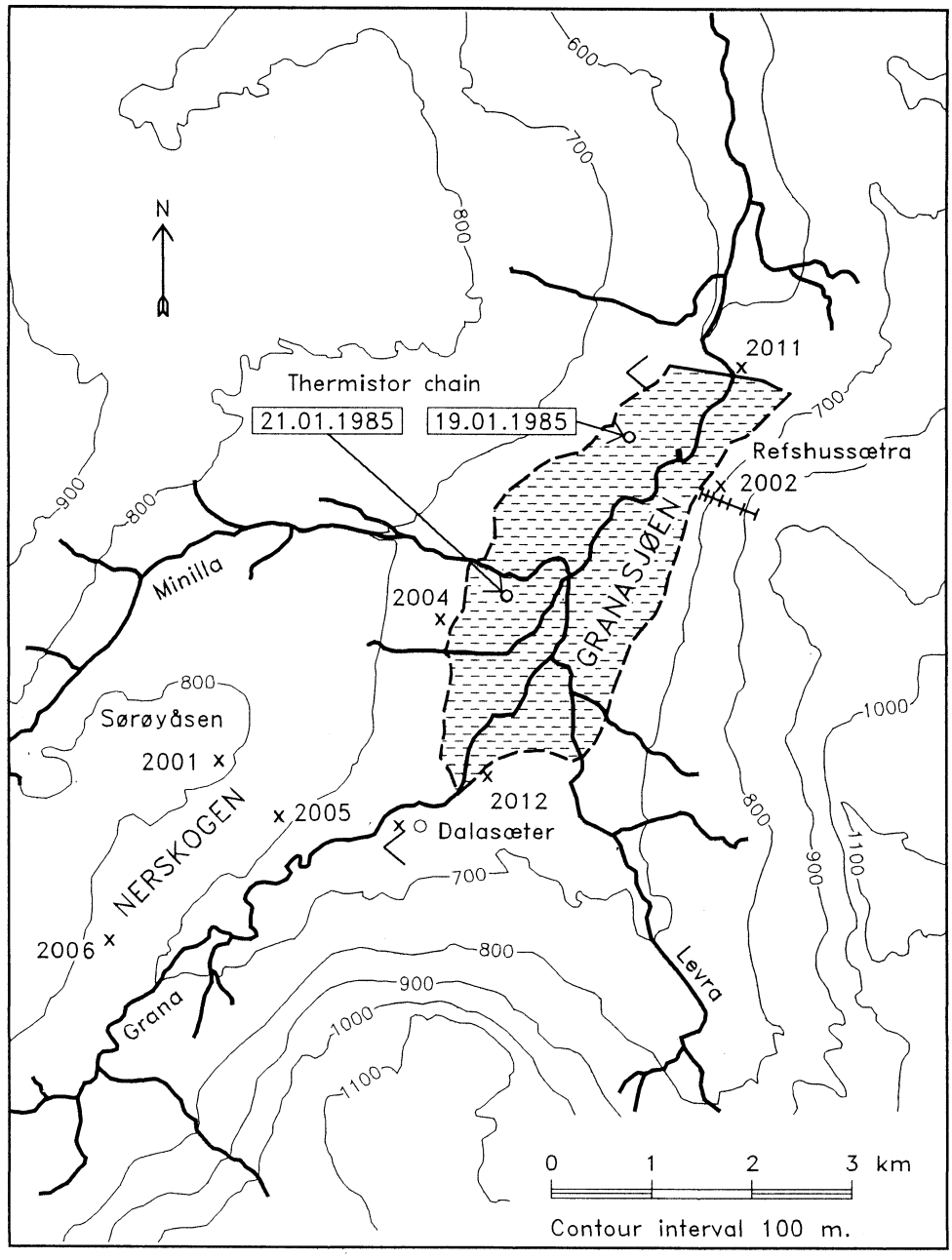


*Figure 3.3. Climatic stations at Lake Granasjøen (cf. Figure 3.20).* ***Before and after regulation****, recording every hour, 2 m above the ground: 2001 Sørøyåsen (800 m a.s.l.), 2002 Refshussætra (7 stations upslope from HRW, 650 m, to 800 m), 2004-2006 Nerskogen I-III (660, 710, and 760 m). Termohygrograph and reading of termometers and anemometer: 2003 Dalasæter (675 m a.s.l.). < photos during snow melt in spring.* ***After regulation only****, recording every half hour: 2011 dam (655 m), 2012 Dalasæter (652 m), sensors at 2 and 10 m level Skaar, 1986).*

In ***June*** the surroundings of Lake Granasjøen were snow-free throughout the project periode. Figures 3.4 and 3.5 depict the cooling and heating effect of the waterbody on ambient air during two days in early June 1984. The largest effects on the air temperature are expected just on such ***radiation days*** (strong insolation during daytime, and great long-wave radiative loss during the night) with warm days and cool nights. In both cases the filling of the reservoir was close to HRW, and the surface temperature was some 9°C, whereas estimated median water level for the first week of June is some 4 m lower (KVO, 1981), and estimated surface temperature 2-3°C lower (Berge et al., 1981).

On June 1, 1984 (Figure 3.4) down-valley wind - from SE - prevailed, and accordingly offshore winds were predominant at both station 2012 and 2002. Differences between these stations in the daytime air temperature 2 m above the ground were small, whereas simultaneous temperatures at station 2011 (dam) in the downwind area were some 2°C lower. On June 4 (Figure 3.5) up-valley wind persisted in daytime, and at station 2012 on the downwind shore, the temperature 2 m above the ground was some 3°C lower than that at 2011 (the dam) most of the day. At station 2002 onshore wind prevailed in the afternoon, 13-17 h, and the maximum relative drop in temperature 2 m above HRW was slightly above 2°C. The mean temperature during this periode was: at the dam 18.5°C , and at Refshussætra (2002) 16.9, 19.0, 19.1, and 19.4°C at 2, 12, 22, and 32 m above HRW respectively. Cool lake air that overruns warm land is obviously quickly heated near the surface. In addition to local solar heating a somewhat sparse wood (mainly birch) gives shelter and induces turbulent mixing of warmer air from above. Thus, the cooling effect of the reservoir on the near surface layer decreases rapidly with increasing distance from the waterbody. At station 2004 (12 m above HRW, some 300 m from the shoreline, and by onshore wind less sheltered by wood than the corresponding sensor at Refshussætra), Skaar found a maximum cooling effect of 0.5-1.0°C in this case. At the stations 2005 (60 m above HRW, 1.5 km from the shoreline), and 2006 (110 m above HRW, 3 km from the shoreline) he found no effects that could be attributed to Lake Granasjøen.

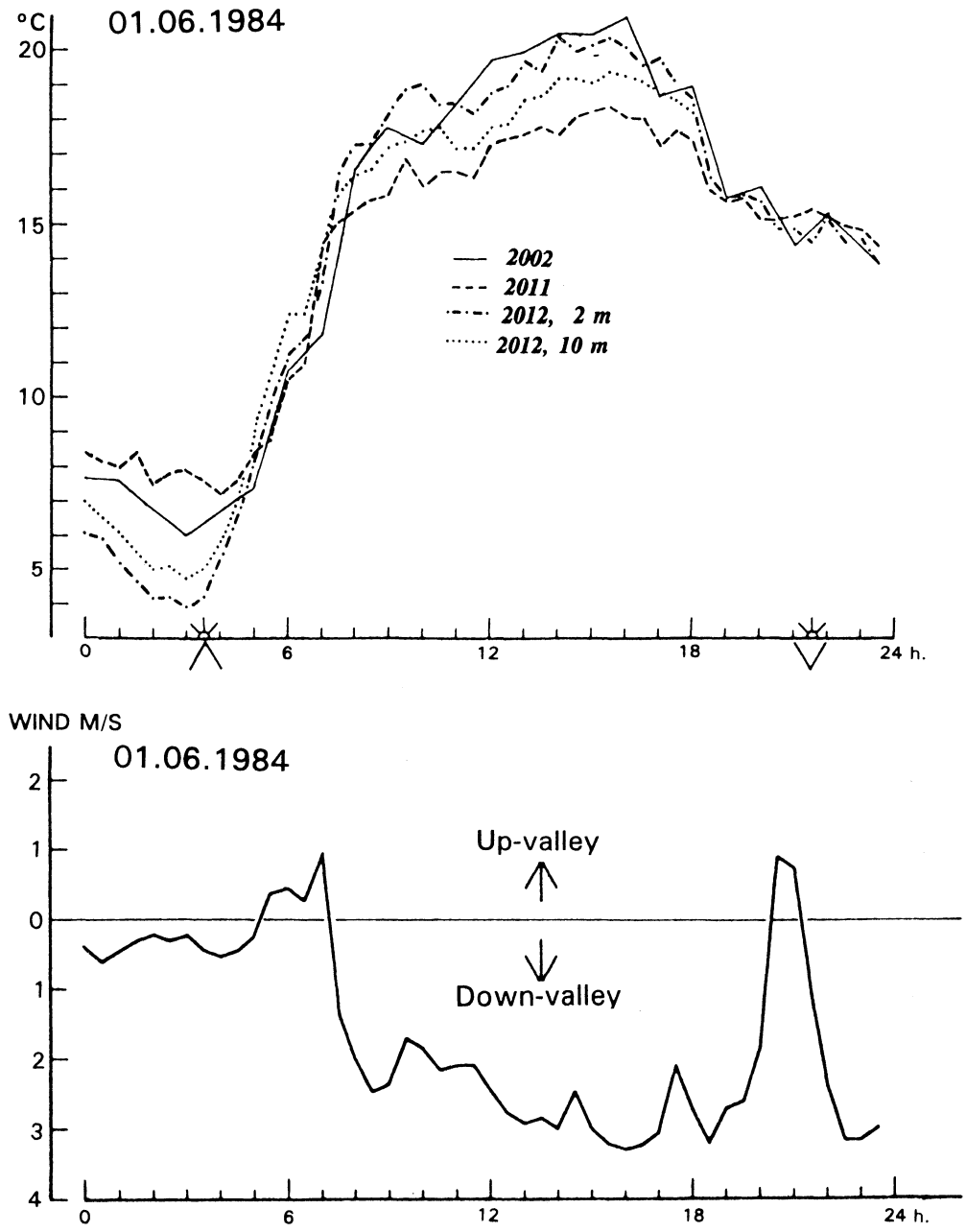


*Figure 3.4. Recordings of air temperature at three stations close to HRW at Lake Granasjøen June 1 1984: 2002 Refshussætra, 2011 dam, and 2012 Dalasæter. Wind recorded at 2012 (Skaar, 1986).*

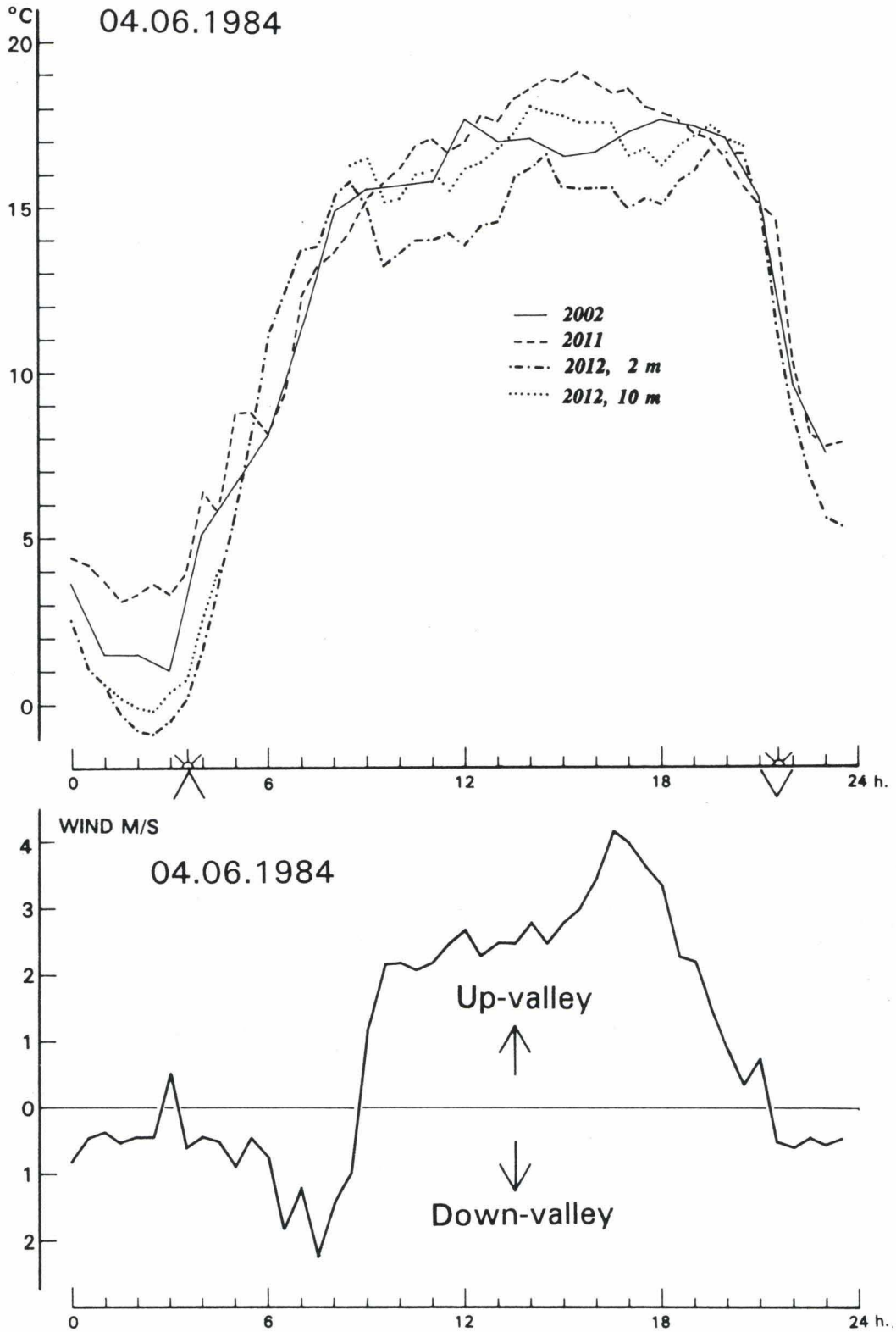


*Figure 3.5. Recordings of air temperature and wind at Lake Granasjøen June 4 1984. Station No as in Figure 3.4 (Skaar, 1986).*

A noticeable feature in these two situations is the interrelation between the tempertures measured 2 and 10 m above the ground at station 2012. By offshore wind in daytime (Figure 3.4) the temperature at 2 m was some 1°C higher than that at 10 m, wheras by onshore wind (Figure 3.5) it was some 2°C lower. The first case clearly illustrates the effect of the solar heating of the ground on clear summer days, and the second case shows that when warm air overruns a cold surface (in this case the temperture difference was up to 10°C) an inversion is formed in the airflow. The latter also partly explains why the cooling effect of a waterbody on air temperature is most marked in a shore zone just above HRW.

During the night a weak down-valley wind persisted in both cases, and the heating effect of the reservoir raised the air temperature some 2°C, 2 m above HRW at Refshussætra, and some 4°C at the dam.

During a clear night with light mountain wind in autumn the heating effect of the reservoir was, as shown in Figure 3.6 even greater; some 2°C at Refshussætra, and some 6°C at the dam. The lake was filled to HRW and the surface tempereature was some 10°C. During the warmest part of the day the air temperature was some 9°C at the dam. As might be predicted, the onshore air flow had no noticeable effect on daytime air temperatures at any station.

To asses average climatic effects of Lake Granasjøen on ambient air temperature the difference method (Appendix 6.1) was applied on data collected at Refshussætra (2002, 7 locations), at the 3 stations (2004, 2005, and 2006) at Nerskogen, and at the reference station Sørøyåsen (2001). The main results for the stations on the slope at Refshussætra are given in a simplified form in Table 3.1, where $\bar{E}_S$ is the mean effect of the regulation on temperatures at Refshussætra, and $\Delta\bar{T} = \bar{T}_R - \bar{T}_S$ is mean difference between Refshussætra and Sørøyåsen. Apparently, there were no significant effects at locations higher than some 30 m above HRW, and subsequently results for those locations are not given in Table 3.1. A positive $\bar{E}_S$ indicates a rise in the temperature due to regulation, and a negative $\bar{E}_S$ indicates a drop.

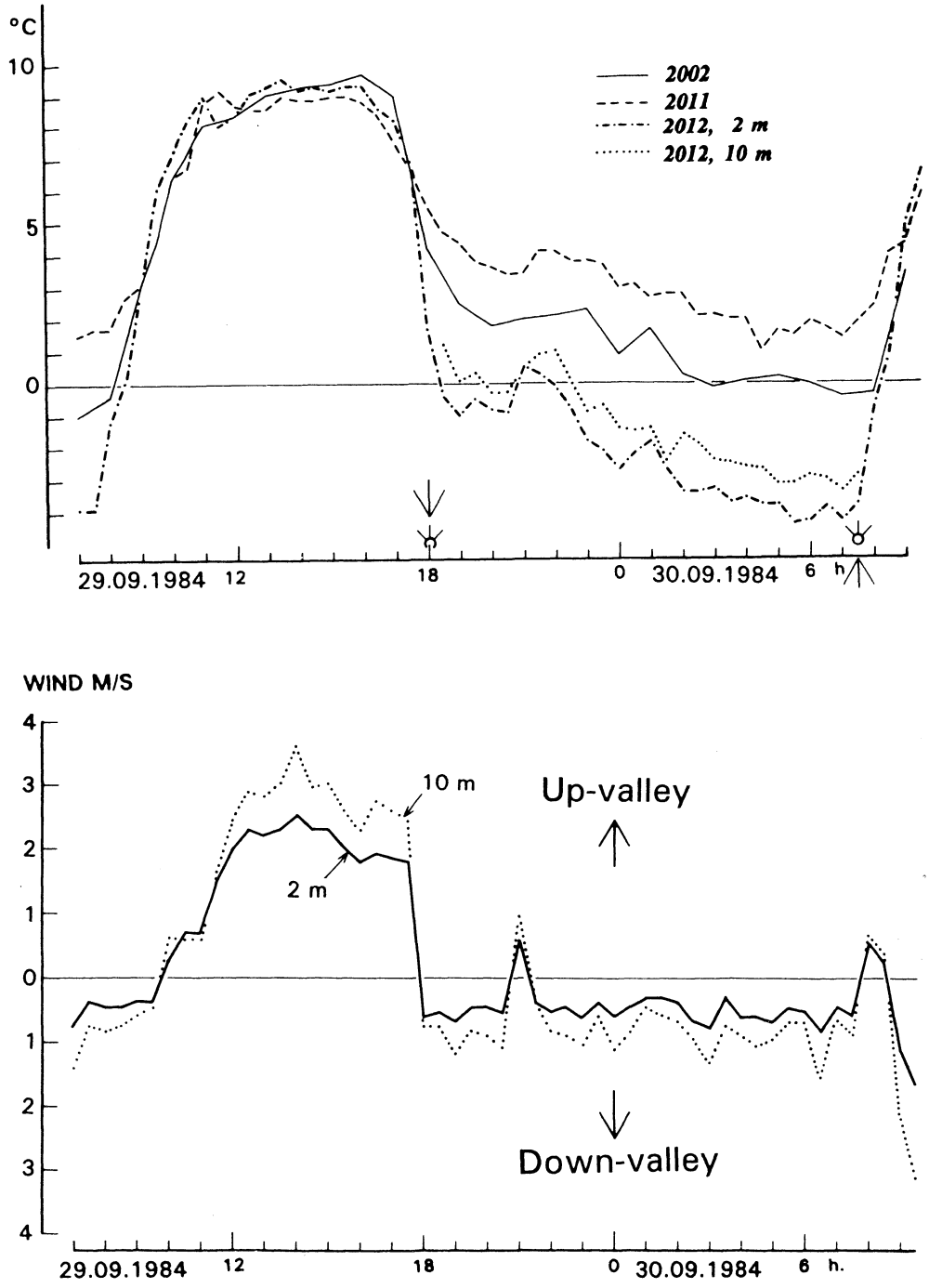


*Figure 3.6. Recordings of air temperature and wind at Lake Granasjøen September 29-30 1984. Station numbers as in Figure 3.4 (Skaar, 1986).*

*Table 3.1. Mean differences, $\bar{E}_S = \Delta\bar{T}_{after} - \Delta\bar{T}_{before}$ , of temperature differences between 2002 Refshussætra (four hights above HRW) and 2001 Sørøyåsen, for the months June, July, August, and September,* ***before (1979-81) and after (1982-83)****. T > 15°C are all cases with air temperature above 15°C at Sørøyåsen, and $T_X$, $T_N$, $T_D$ are maximum, minimum, and diurnal mean temperatures respectively.*

| Hight above HRW m | 2 | 12 | 22 | 32 | 2 | 12 | 22 | 32 |
|---|---|---|---|---|---|---|---|---|
| Month | | | June | | | | July | |
| T > 15°C | -1.0 | -0.3 | -0.2 | 0 | -0.6 | -0.3 | -0.2 | 0 |
| $T_X$ | -1.0 | -0.4 | -0.2 | 0 | -0.6 | -0,4 | -0.2 | 0 |
| $T_N$ | 0.5 | 0.2 | 0 | 0 | 0.3 | 0.2 | 0 | 0 |
| $T_D$ | -0.3 | -0.1 | -0.1 | 0 | -0.3 | -0.1 | 0 | 0 |
| Month | | | August | | | | September | |
| T > 15°C | -0.6 | -0.3 | -0.2 | 0 | 0 | 0 | 0 | 0 |
| $T_X$ | -0.4 | -0.2 | -0.2 | 0 | 0 | 0 | 0 | 0 |
| $T_N$ | 0.2 | 0.1 | 0 | 0 | 0.9 | 0.5 | 0.4 | 0.1 |
| $T_D$ | -0.2 | -0.1 | 0 | 0 | 0.5 | 0.4 | 0.1 | 0.1 |

It is noted that the mean cooling effect in the shore zone at high temperatures in daytime is most pronounced in June. It is marked also in July and August, whereas in September there is apparently no cooling. The heating effect at night ($T_N$) which is also greater in June than in July and August, is most pronounced in September. The effect on diurnal mean temperature is not statistically significant for any of the months June-August. Nevertheless, the results indicate that the regulation has caused a small lowering of the mean temperature in the shore zone during the summer months. Just above HRW the drop is probably 0.2-0.3°C, and the effect obviously vanishes some 15-20 m above HRW. In September there is a statistically significant rise of some 0.5°C in the mean temperature just above HRW.

### *3.2.2. Winter at Lake Granasjøen*

The expected effect of Lake Granasjøen - filled to HRW and frozen up - on the down-valley cold air flow in calm, clear weather in winter is schematically depicted in Figure 3.7. Dashed lines indicate thickness, vertical temperature profile, and stream arrows before regulation, and full lines indicate conditions after regulation. Draining of the ice-covered reservoir induces only small changes in the "top level" and the speed characteristics of the flow above HRW. However, the vertical temperature distribution changes in a manner indicated by the dotted curve in Figure 3.7.

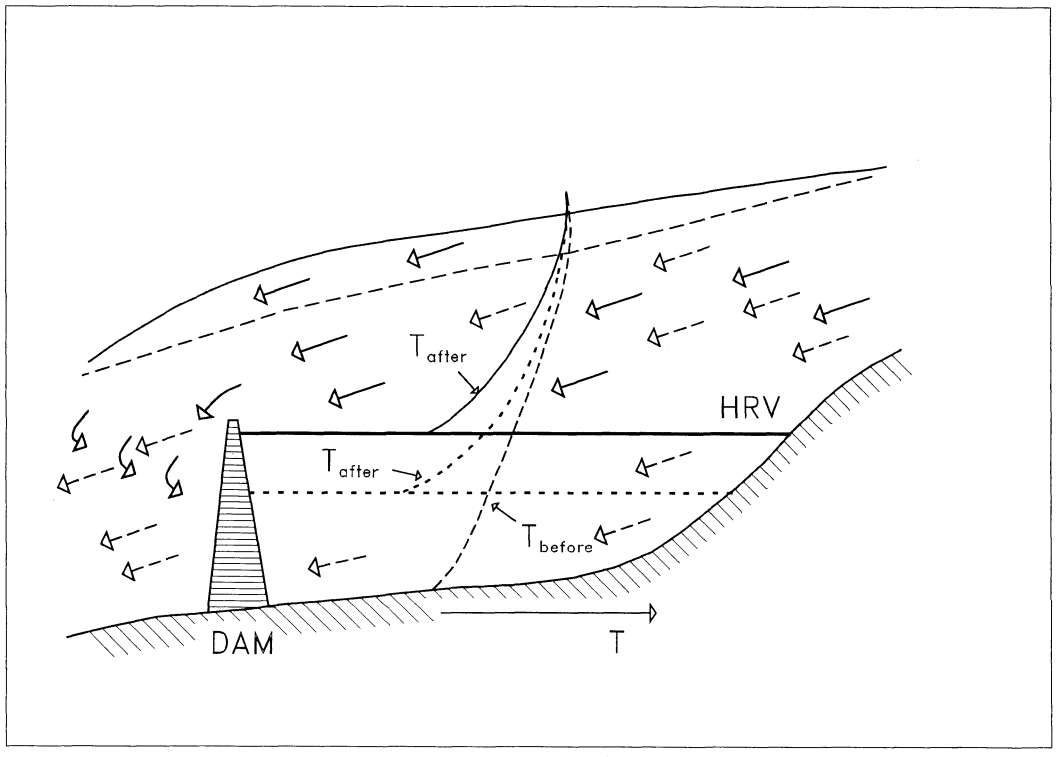


*Figure 3.7. Schematic illustration of cold air flow in the Lake Granasjøen area during calm, clear nights in winter.* ***Dashed curves*** *indicate thickness and vertical temperature distribution in the flow before the construction of the dam,* ***full curves*** *indicate corresponding conditions after regulation with ice-covered reservoir filled to HRW.* ***Dotted curve*** *indicate temperature distribution by ice-covered reservoir drained to a level well below HRW.*

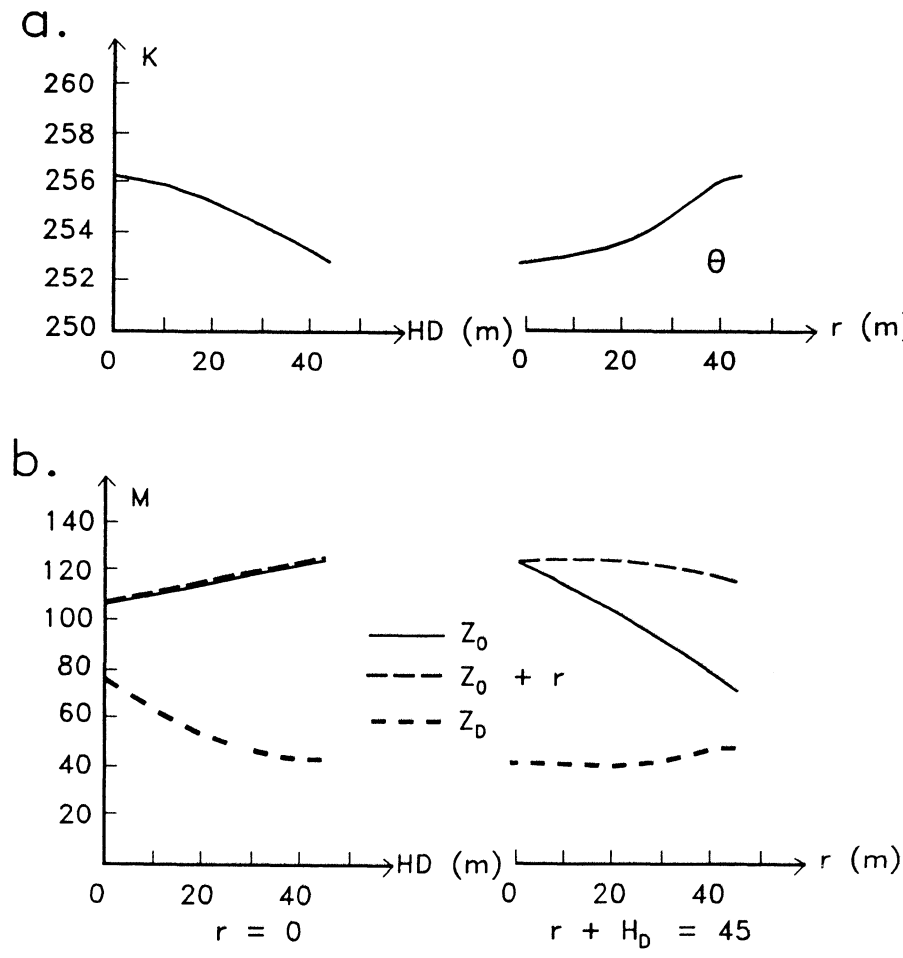


*Figure 3.8. Characteristics of a stationary cold air flow in the Lake Granasjøen area.* ***Left:*** *As a function of dam-height, $H_D$ , but with no water trapped behind the dam (r=0).* ***Right:*** *As a function of water depth, r , of the ice-covered reservoir behind the 45 m high dam. a) Mean potential temperature, $\theta$ , of the inversion layer above the reservoir. b) Thickness of the inversion layer in the reservoir area, $Z_0$ , thickness of the flow on the dam, $Z_D$ , and hight above the valley floor of the inversion layer, $Z_0$ + r (after Andersen and Skaar, 1987).*

The main results of model calculations for steady state drainage situations (Andersen and Skaar, 1987) are summarized in Figure 3.8. To the left in the figure is shown how the characteristics of the cold air flow in the valley changes with increasing hight of the dam when there is no water in the reservoir. As the height of the dam is enlarged from 0 to 45 m the thickness of the flow in the reservoir area increases gradually from circa 108 to 120 m, whereas the thickness of the flow at the dam site decreases from some 72 to 41 m; i.e. the top of the cold air flow is lifted some 14 m.

The mean temperature of the cold air layer above the reservoir decreases with increasing hight of the dam, because the coldest air is trapped behind the dam, while warmer air flows over. To the right in Figure 3.8 are depicted the changes in these conditions when the full, ice-covered reservoir is gradually drained. It is noted that the tickness of the cold air layer above the central part of the reservoir increases from circa 72 to 126 m, whereas simultaneously the thickness over the dam decreases from circa 49 to 41 m. This indicates that by full resevoir the top of the cold air flow is slightly lower above the reservoir and somewhat higher over the dam than by completely drained reservoir.

In the model it is assumed that the mean temperature of the cold air layer above the full reservoir is equal to corresponding temperature in this area before the construction of the dam.

Stationary flows and stable pooles of cold air are rare phenomena. The drainage of cold air downslope or down-valley often occurs as intermittent surges rather than a continuous flow. This may result in rapid temperature fluctuations. An example of recordings in a pool of cold air at Lake Granasjøen by means of a vertical chain of 10 thermistors lifted by a balloon (cf. Figure 3.1) is given in Figure 3.9. The sampling rate was once per minute. Calm down-valley winds prevailed. Half hourly mean wind speeds were 0.5 - 1.0 m $s^{-1}$ within the pool, and some 2 m $s^{-1}$ on the dam, with gusts about twice these values. The somewhat chaotic temperature fluctuations clearly reveal the gustiness which probably may be attributed to effects

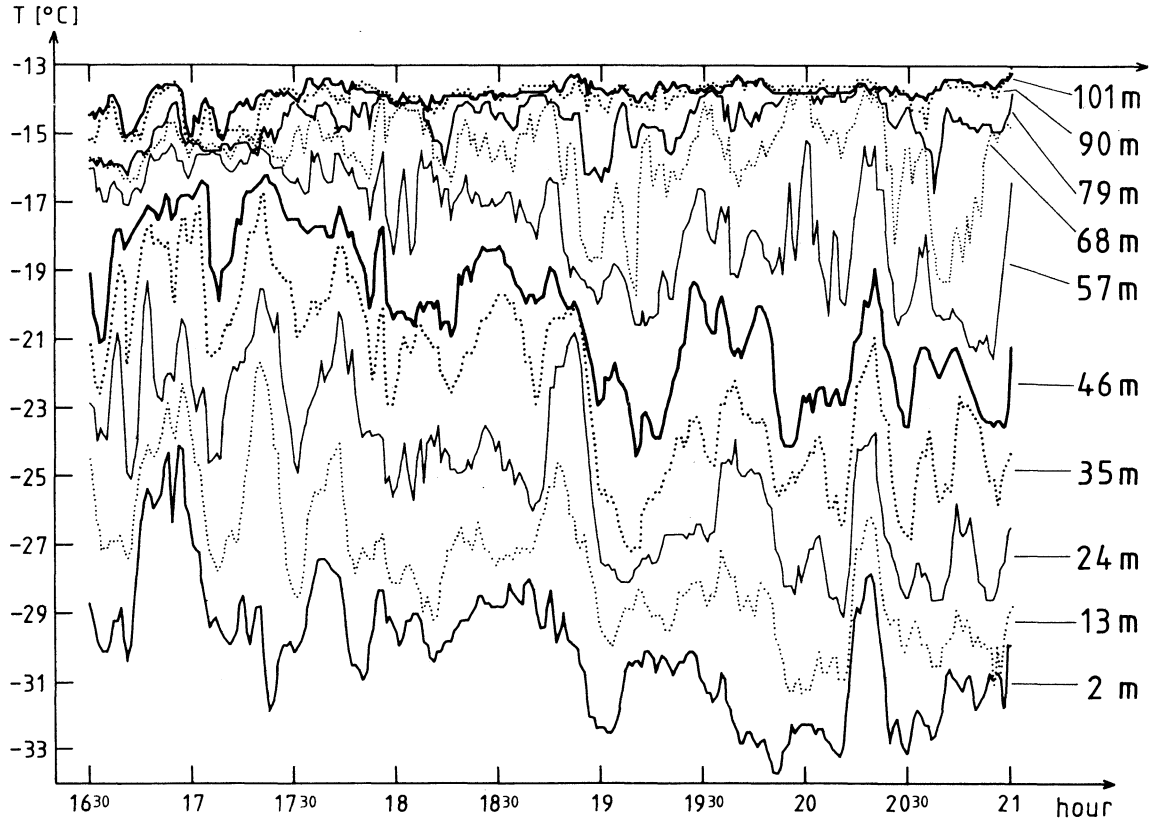


*Figure 3.9. Temperature recordings made at 10 levels above the snow-covered ice of Lake Granasjøen (cf. Figure 3.3) in calm, clear weather 19.01.85, 1630- 2100 CET. The reservoir was drained to 12 m below HRW, i.e. 17 m below the top of the dam (after Andersen and Skaar, 1987).*

of surges of cold air, internal waves, and the weak gradient wind aloft. Temperature changes greater than 5°C occured within a few minutes, and the fluctuations deacreased towards the top of the inversion layer, which in this case had a thickness of 90-100 m.

Figure 3.10a shows that in this situation hourly recordings of air temperature on the slope at Refshussætra deviated strongly from those of the thermistor chain above the reservoir. The mean vertical profiles for four hours (Figure 3.10b) are smoother, but the differences are marked. At the lower levels temperatures recorded

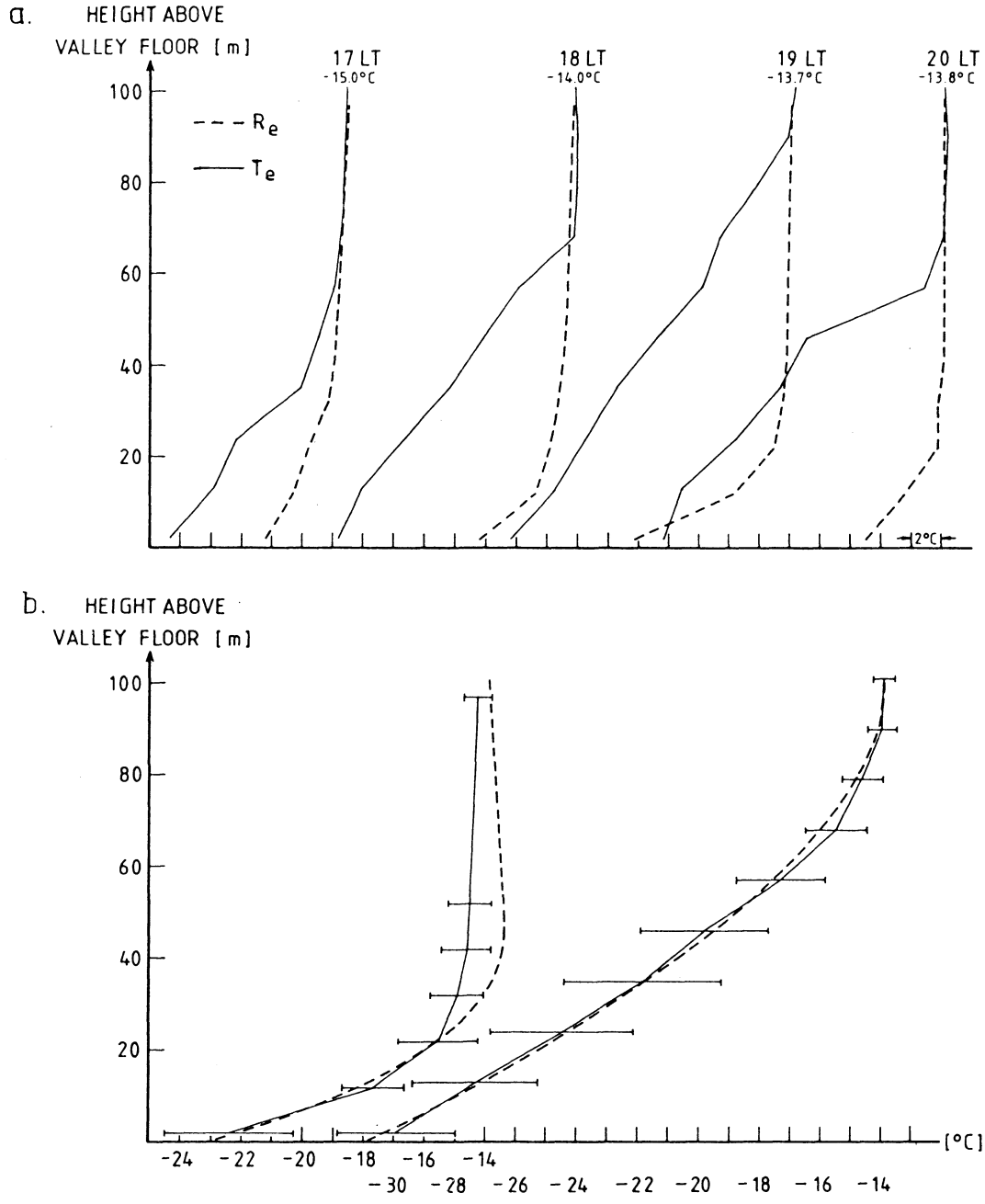


*Figure 3.10. a) Hourly vertical temperature profiles recorded on the slope at Refshussætra, Re , and in the thermistor chain above the reservoir, Te , 19.01.85 (cf. Figure 3.3), instantaneous values. b)* ***Full lines:*** *average recorded profiles at Refshussætra (1600-2000 CET) to the left, and in the thermistor chain (1630- 2030 CET) to the right.* ***Dashed lines:*** *theoretical profiles (power law) as found by least square fits to the observations (after Andersen and Skaar, 1987).*

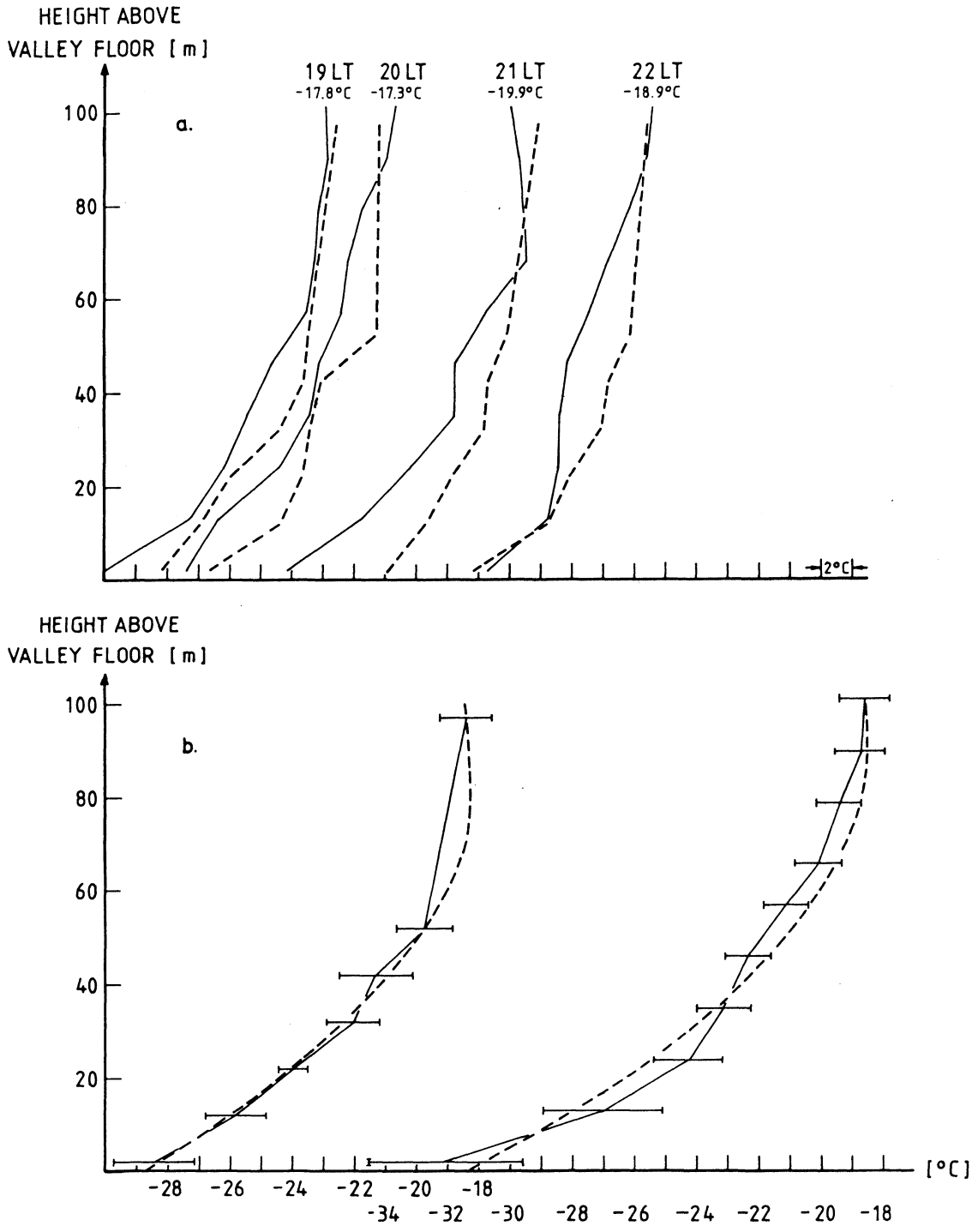


*Figure 3.11. As Figure 3.10, but for 21.01.85, and for thermistor chain recordings made in the interior part of the reservoir (cf. Figure 3.3). a) Instantaneous values. b) Averages 1900-2300 CET (after Andersen and Skaar, 1987).*

by the thermistor chain are much lower than those recorded on the slope, and there is a strong inversion up to a hight of 80-90 m above the valley floor, whereas an inversion of corresponding strenght on the slope appears at levels below some 20-30 m. Under similar conditions - calm down-valley winds and strong temperature fluctuations in the pool of cold air - on 21 January 1985, the thermistor chain was situated further away from the dam (cf. Figure 3.3). Figure 3.11 shows that in this case the temperature recordings in the two profiles match fairly well. The reason why the top of the pool of cold air appears higher up the slope at Refshussætra in this case than in the case depicted in Figure 3.10, may be the difference in the direction of the gentle gradient wind.

On 21 January the SSW-wind aloft was directed along the axis of the valley, whereas the NE-wind aloft on 19 January was directed slightly cross-valley, which may have pushed the flow of cold air somewhat away from the eastern slope of Lake Granasjøen. The results presented in Figures 3.9, 3.10, and 3.11 clearly illustrate the problem of getting representative measurements of the temperature field in a flow (pool) of cold air, and that the assumption of stationary conditions can be rather unrealistic.

Since corresponding measurements were not made before the construction of the dam, its possible effect on the gustiness of the flow of cold air cannot be assessed in any detail. The hourly recordings on the slope at Refshussætra, and the thermogaph recordings at Dalasæter indicate that on calm, clear winter days marked oscillations within the flow of cold air occurred quite frequently also before regulation. However, the oscillations above the smooth surface of the ice-covered reservoir are in most cases evidently stronger, and appear somewhat higher upslope than in the virgin valley before the dam was built.

To asses the effects of the dam on the temperature characteristics of the cold air flow an anlysis of the data from Refshussætra for the winters 1979/80-1984/85 has been made (Utaaker and Skaar, unpublished). When calculating the difference ***before*** - ***after*** regu-

lation for selected cases of calm, clear weather, they applied four-hourly averages in order to smoothen the effects of the gustiness of the flow on internal temperatures. It was also presumed that at 800 m a.s.l. (150 m above HRW) at Refshussætra the temperature was not affected by the construction of the dam.

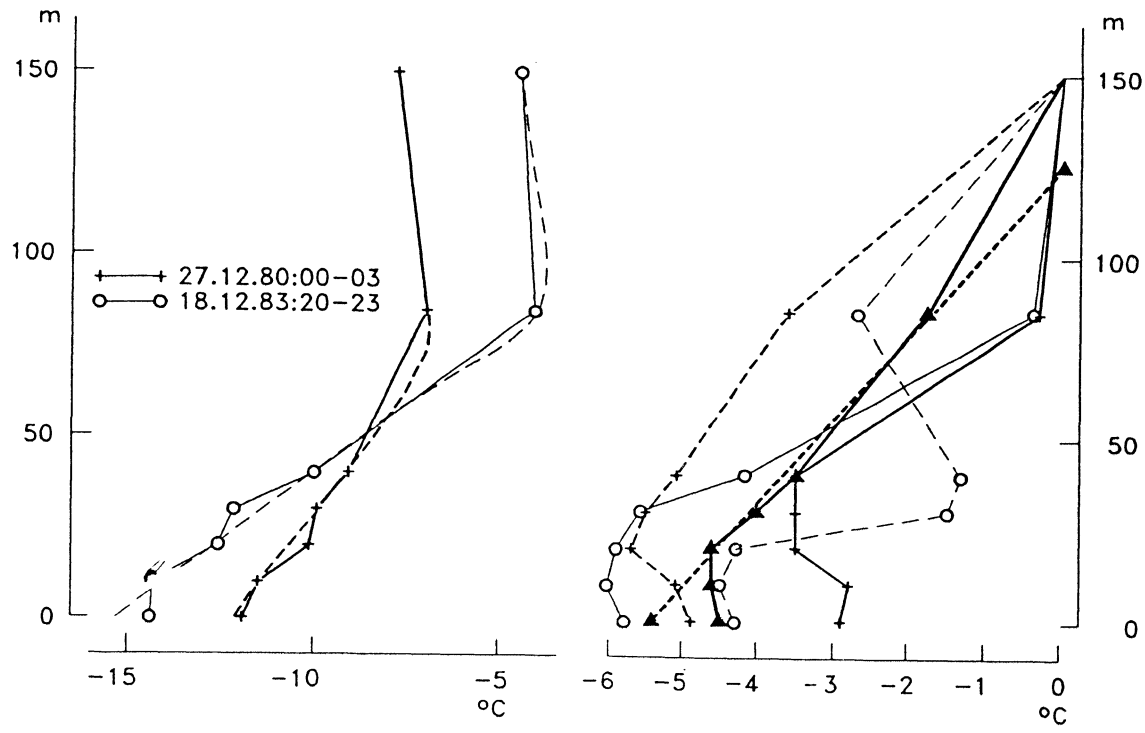


*Figure 3.12. Four-hourly averages of temperature recordings in the profile at Refshussætra on calm, clear nights in December, by snow-covered ground, and, after regulation, also ice-covered reservoir filled to some 4,5 m below HRW.* ***Left:*** *One example* ***before*** *and one* ***after.*** ***Right:*** *Four examples of differences,* ***before-after.*** *The full line marked by triangles depicts the mean differences of the four examples, and the straight, dashed line indicates a linear reduction in the effect of the regulation with height (Utaaker and Skaar, unpublished).*

The examples in Figure 3.12, from December with reservoir filling some 4.5 m below HRW, indicate that the influence of the regulation on ambient air temperature is largest in the shore zone up to 20-30 m above HRW, and that it diminishes further upslope. The cooling effect appears to be slightly stronger at 22 m above HRW than at the two lower points of measuring. The reason may be a stronger tendency to cold air stagnation at the former site than at the two latter. Neither can small systematic errors in the temperature sensors be excluded. However, the variation in the vertical profiles within the inversion layer - as measured along the slope - and fairly large ranges in the four-hourly mean temperatures at all levels indicate considerable "noise" caused by the gustiness of the flow of cold air.

Nevertheless, the averages of the 4 four-hourly means (Figure 3.12, full line marked by triangles) for the levels 22-87 m above HRW form a nearly straight line. This indicates that in well developed inversions on calm, clear winter nights the cooling effect of the resevoir decreases nearly linearly with height above HRW. Consequently the top of the "mean" inversion in Figure 3.12 would reach a height of some 125 m above HRW, i.e. markedly heigher than expected from the model calculations and the measurements depicted in Figures 3.10 and 3.11.

Analogous differences for January with reservoir filling some 12 m below HRW showed a pattern akin to that of Figure 3.12, but the mean differences were markedly smaller (1.5-3,5°C in the shore zone). In the winters ***after*** the construction of the dam very few situations favouring development of strong inversions occured during February and March. However, the preliminary, uncertain results indicate that the effect of the dam on air temperatures above HRW decreased with increasing drainage of the reservoir, and no apparent effects were measured at higher levels than 40 m above HRW during March-April.

Based on the results of the recordings at Refshusssætra and on model calculations the probable effects of Lake Granasjøen on ambi-

ent air temperature on calm, clear winter nights are summarized in a simplified way in Table 3.2. It is emphazised that the numbers in this table should be considered an illustration of expected effects rather than exact results. By moderate or strong winds, and by cloudy weather there are obviously no measurable effects of the dam on ambient air tempeature.

*Table 3.2. Probable lowering of minimum temperatures (°C) caused by the dam during calm, clear winter nights, at different levels above HRW around the ice-coverd Lake Granasjøen, and at different filling of the reservoir.*

| Hight above HRW m | 2 | 12 | 22 | 32 | 42 | 87 |
|---|---|---|---|---|---|---|
| Filled to HRW | 5.5 | 5.0 | 4.5 | 4.0 | 3.5 | 1.5 |
| 12 m below HRW | 2.8 | 2.6 | 2.4 | 2.2 | 2.0 | 1.1 |
| Drained to LRW | 1.0 | 0.8 | 0.6 | 0.4 | 0.2 | 0 |

Nordli (1994) studied the effect on ambient climate of ***Lake Blåsjøen,* situated** on a rugged, high plain, Ryfylkeheiane (circa 59° 20'N, 7°E). This main reservoir of the Ulla-Førre regulation is the largest artificial lake in Norway. When filled to HRW (1055 m a.s.l.), the water surface is 88 $km^2$. Before regulation several smaller lakes covered some 63 $km^2$ of this area (Figure 3.13).

During the period 1975-1993 (***before, under, and after*** the construction of the dams) temperature and humidity of the air, wind speed, and wind direction were recorded at the stations Sanddokkryggen (50 m above, and at a distance some 250 m from HRW) and Høgaloft (38 m above, some 100 m from HRW), and temperature was recorded at the secondary stations Sanddokki (5 m above, 10 m from HRW) and Høgaloftkvelven (23 m above, 40 m from HRW). Measuremets of water surface temperature at different locations after 1986, when filling had linked the different parts into one reservoir, showed negligible differences between measuring points. Thereafter the water level has remained above 1040 m a.s.l.

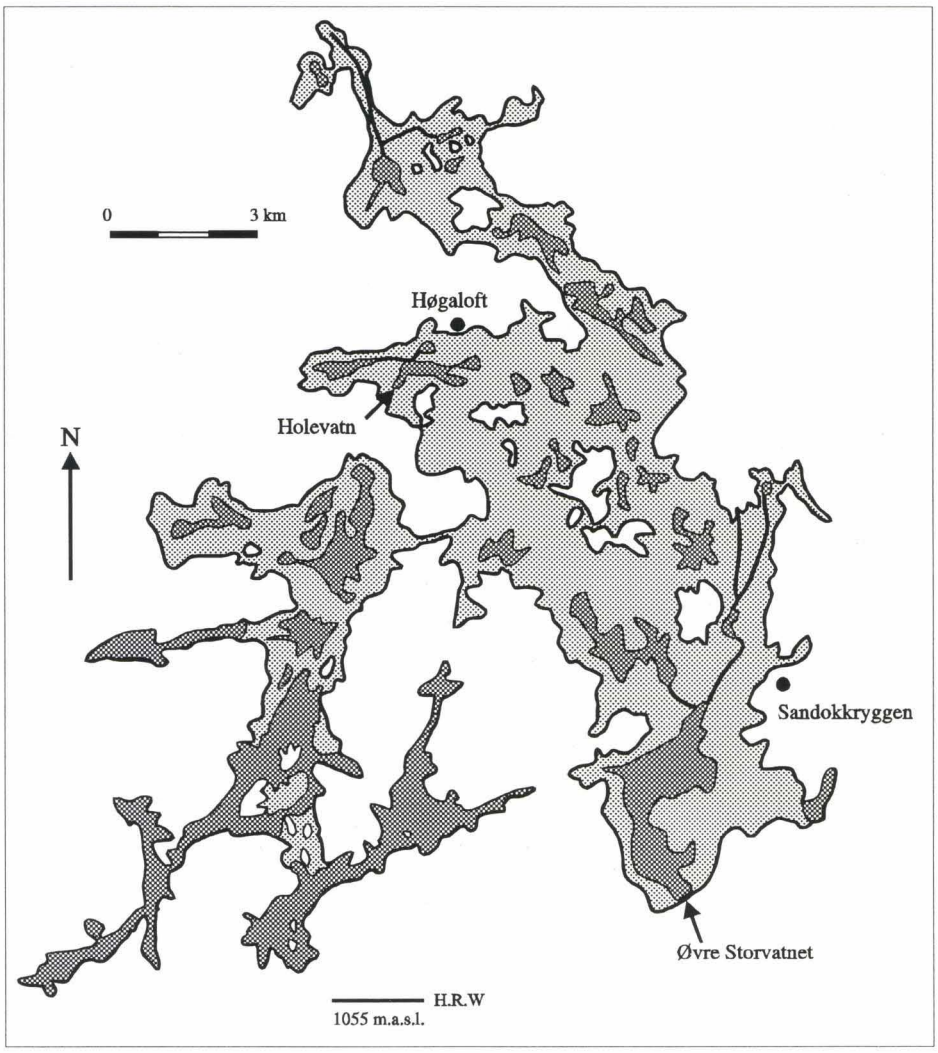


*Figure 3.13. Map of Lake Blåsjøen at HRW. Darkly shaded areas depict lakes prior to regulation.* ***Above:*** *The reservoir as seen from Høgaloft; water level 1049 m a.s.l. (after Nordli, 1994).*

Before regulation the water level was always below 976 m a.s.l. in Lake Holevatn and Lake Øvre Storvatn, the lakes closest to Høgaloft and Sanddokki respectively.

During spring-summer water surface temperature in general stayed lower than ambient diurnal air temperature into August. In September the water was usually warmer than ambient air, and in October average water temperature was always higher than that of environmental air. Usually, some ice formed in November, but time of freeze-up varied between late November and late December at Høgaloft, and between late November and early February at Sanddokki.

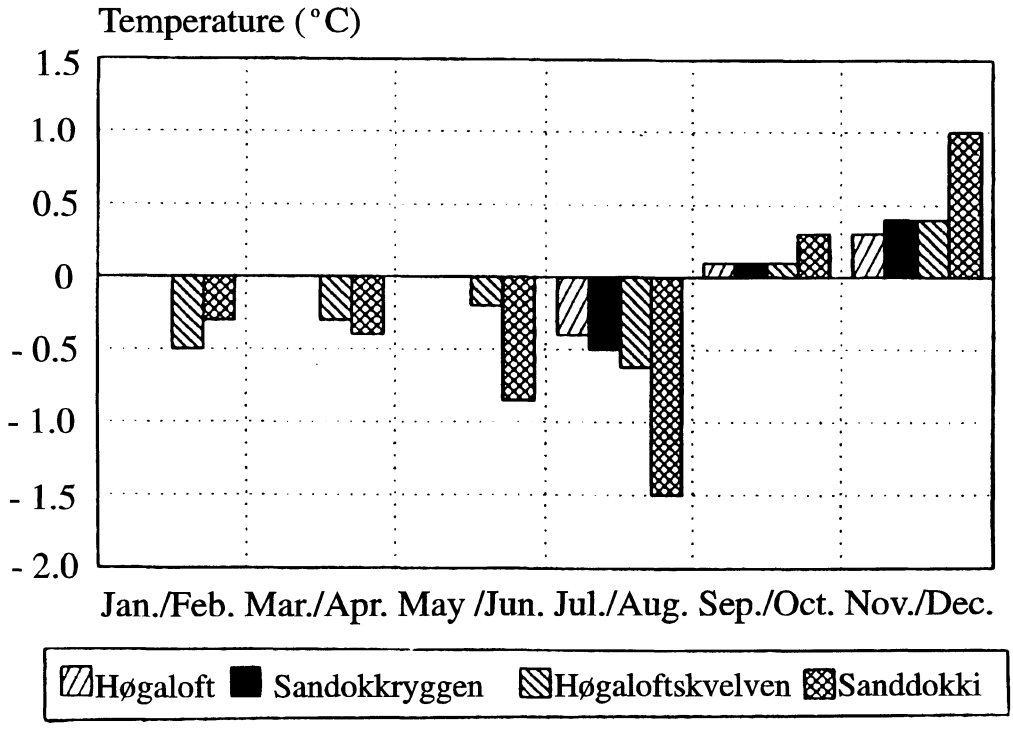


*Figure 3.14. Changes due to regulation* in avearge air temperature at the four measuring sites at Lake Blåsjøen. Differences between recordings after regulation and before regulation for successive two-month periods (after Nordli, 1994).

Figure 3.14 depicts effects of the regulation on average air temperature at the four stations for successive two-month periods. Testing the results, Nordli concluded that possible errors in these averages are ±0,3°C.

Briefly summarized the results in Figure 3.14 show:

***January/February and March/April.*** The regulation has caused a significant drop in average air temperture at the two lower stations. The explanation is that in inversion situations the dams impede the outflow of accumulated cold air over the (normally) ice-covered reservoir. Moreover, the base of the local ground inversion is closer to the temperature stations than before regulation (cf. also Figure 3.7). The apparent greater effect at Høgaloftkvelven (23 m above HRW) than at Sanddokki (5 m above HRW) in January/ February may be due to occasions of ice-free water near the latter location while the lake was frozen up near the former. However, this difference is not statistically significant. No measurable changes in average air temperature at the two higher stations evidence that inversions developing above the reservoir in this wind-swept terrain are fairly shallow, and rarely extend above surrounding knolls higher than some 30-40 m above HRW.

***May/June.*** Before regulation snowmelt resulted gradually in wide areas of snow-free ground within the reservoir area, which after regulation is covered by snow/ice, or water close to 0°C. This has changed drastically local solar heating of the ground. Therefore, the the vast, cold surface of the reservoir induces a significant lowering of ambient air temperature. As expected the drop in average temperature was most marked at the measuring site closest to the lake, and the difference between temperature drops at the levels 5 and 23 m above HRW was more marked than in March/April.

In ***July/August*** average water surface temperature of the reservoir was - as stated above - in general higher than ambient, average air temperature. Lake air, cooled over the water, invading surrounding

land resulted in a drop of some 1.5°C in average air temperature at Sandokki due to regulation. Smaller but significant effects were recorded also at the three higher sites in greater distances from the cold water surface.

In ***September/October*** higher average water than air temperature raised ambient air temperature (which were below long-time averages for the season) slightly, but significantly. As expected, the effect was greatest at the site closest to the water.

During ***November/December*** the reservoir normally froze up. After filling above 1040 m a.s.l. in 1986, open water was more frequent than ice in the Sanddokk area, but this was hardly the case in the Høgaloft area. Compared to results from September/October this difference in ice conditions is evidenced clearly in the considerably greater rise in average air temperature at Sanddokki than at Høgaloftkvelven due to regulation. The regulation also seems to have raised average temperature at the two higher measuring sites. According to Nordli these differences are not statistically significant. However, model calculations indicated significant temperature changes due to regulation also at those levels.

Nordli (op. cit.) also tested the effect of ice-covered versus open water during the months October-February 1986/87-1992/93, i.e. after regulation when the water level was always above 1040 m a.s.l. He found that average diurnal air temperature at Sanddokki (5 m above, and 10 m from HRW) was some 1°C higher by open than by ice-covered reservoir. The difference decreased to some 0.5°C at Høgaloftkvelven (23 m above, and 40 m from HRW). At the measuring sites 38 and 50 m above, and at distances 100 and 250 m from HRW respectively, corresponding values were some 0.4°C.

## *3.3. Along watercources with reduced discharge*

In Norway the development of hydropower usually reduces the flow in the waterways between the reservoir(s) and the outlet of the power station strongly, espcially in spring - summer - autumn. The outcome is lower level and reduced area of water, and changes in the temperature and ice conditions of the watercourse.

### *3.3.1. Consequences of lower water level*

The water level in a river may affect the groundwater conditions in adjacent soils. Especially on lowlying land along a river, a reduction in the flow of water can lower the groundwater table, which may influence the water supply to the plants (Myhr, 1970, 1971, and 1982). Reduced water supply from the groundwater - especially during dry periods - results in less available water to plants in the root zone. This may impede evapotranporation, and possibly cause crop failure (op. cit.). Some of the available radiant energy that under more favorable soil moisture conditions would have been used for evapotranspiration (latent heat), must be dissipated as sensible heat which causes additionel heating of the vegetation, the upper soil, and the air near the surface. Drying reduces both the thermal conductivity and the heat capacity of the soil. Particularly on bare soil or in sparse vegetation the surface temperature will be higher on sunny days, and the heat storage lower in dry than in moist soil. In a radiation night (calm, clear weather) succeeding such a day, the surface temperature will usually be lowest on the dry soil.

At Sletten Farm in Skjåk extensive investigations on environmental effects of changes in the water level in the River Otta were conducted during the summers 1975-81. Gislefoss (1980) and Gislefoss and Skaar (1981) studied effects on groundwater, soil moisture, and micro-climate on adjacent land, and Myhr (1982) assessed crop yield (Figure 3.15).

*Figure 3.15. The experimental area at Sletten in Skjåk as seen towards SW (Photo E. Skaar).*

The experimental area (Figure 3.15) was divided in two equal parts, planted with grass and barley respectively. Each part was then divided in two, of which one was irrigated whenever the ***soil moisture tension*** in the root zone rose above 0.4 bar ($x10^6$ Pa), and the other received natural rainfall only. The groundwater level was recorded in 3 points by means of limnigraphs, and measured dayly in 7 more points. At the depth of 15 cm tension was measured daily in 36 points. Hourly recordings of air and soil temperatures were made in 5 points in a transect normal to the river bank.

Figure 3.16 depicts the result of some micro-meteorological measurements above nonirrigated grass in the summer (May-August) of 1977. During this period rainfall, 98 mm, was somewhat below normal, 125 mm, and air temperature was near normal. It should be noted that average precipitation in Skjåk is among the lowest in

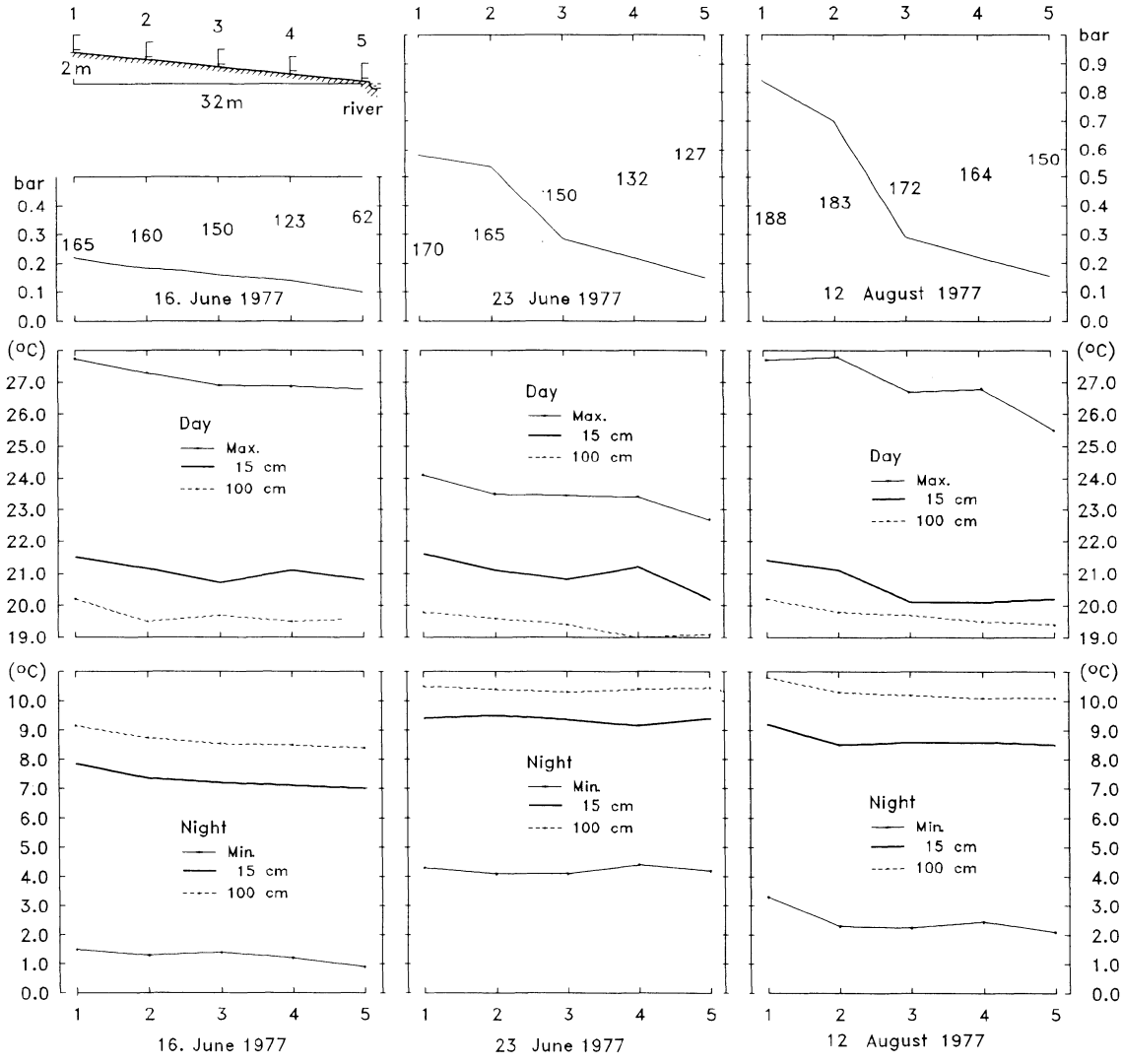


*Figure 3.16. Variation in different elements at the test field at Sletten in Skjåk (section: upper corner left) for 3 selected summer days, 1977.* ***Upper row:*** *Groundwater level (cm below the surface in numbers), and tension (bar) at some 15 cm depth in the root zone.* ***Lower rows:*** *Maximum, average daytime, and nighttime, and minimum temperature (°C) 15 cm, and average day- and nighttime temperature 100 cm above the ground (after Gislefoss and Skaar, 1980).*

Norway, and average, estimated ***potential precipitation deficit (potential evapotranspiration - precipitation)*** at the end of August is 150-200 mm. Irrigation in agriculture has been practised for centuries in this district. The interval from ***field capacity*** (0.1 bar) to 0.5 bar tension represents optimal availability of water for plant use. By desication of the soil beyond 0.5 bar plants may suffer draught stress, and irrigation is advised whenever this threshold is passed (Myhr, 1982). This implies that on 16 June available water for plants was optimal at all 5 measuring points (Figure 3.16). In points 3-5 soil moisture conditions were optimal also on 23 June, and on 12 August, whereas in the points 1 and 2, further away from the river, there was moderate draught stress on 23 June, and severe draught stress on 12 August.

It is noted that air temperatures at the levels 15 and 100 cm did not vary in step along the section. In daytime temperatures 15 cm above the ground were from 0.4 to 2.2 °C higher than corresponding temperatures 100 cm above the ground, whereas at night they were from 0.8 to 1.8°C lower at the lower than at the higher level. The variations in the maximum and the minimum temperatures were mostly more pronounced than those of the average daytime and nighttime temperatures.

Evidently, the moderate draught stresss on 23 June in points 1 and 2 did not affect any of the recorded temperatures. However, the severe draught stress on 12 August caused apparently a "local" rise of some 0.5°C in daytime temperatures, and somewhat more in the maximum temperatures at the 15 cm level, whereas no significant effects were recorded at the 100 cm level. Neither the minimum nor the average nighttime temperatures seemed to be affected on 12 August. Obviously, soil moisture content in the root zone played a minor role in the daytime storage, and nighttime release of heat, probably because a fairly dense vegetation formed an effective barrier to heat transfer to and from the subsoil.

The examples in Figure 3.16 give the main results of the effect of changes in the water level in the river (the groundwater level) on

the air temperature above the nonirrigated test areas. Also at the depth 5 cm in the soil there was a tendency to higher daytime temperatures under high stress conditions, but the effect was less apparent than at the 15 cm level above the ground.

If lowering of the water level drains adjacent bogs or marshes, the induced changes in ambient temperature may be markedly larger than those recorded at Skjåk. However, also here the changes will be restricted to the near surface air and soil.

### *3.3.2. Consequences of changes in water temperature and water area*

When the water flow in a river is reduced during the ice-free season, the water temperature usually tends to vary stronger in step with ambient air temperature, and diurnal amplitudes increase. The seasonal heating in spring, and cooling in autumn is more rapid, and temperatures in summer are in general higher than prior to regulation.

Nordli (1988) studied the effect of the regulation of Ulla River on air temperature at 3 altitudes (64, 85, and 113 m a.s.l.) at Ulladal-Gil some 4 km up-valley from the fjord (Figure 3.17). His main conclusions on the effects of the regulation were:

1. During the months ***January-April*** comparatively small changes in discharge and water temperature have not affected local climate measurably.

2. During the months ***May-July*** the huge disharge of cold meltwater prior to regulation is considerably reduced. This has caused a significant rise in average air temperature at the valley floor; some 0.6 ± 0.4°C at the station on the brink of the river, 64 m a.s.l. A temperature rise of 0.1-0.3°C - not statistically significant - is indicated also at the station 21 m above the valley floor.

3. Data for the months ***August-September*** were inadequate for a statistical analysis. However, Nordli anticipates effects similar to those found for October-December (see below).

4. Also during the months ***October-December*** regulation has reduced discharge considerably, but changes in water temperature are less than in spring and summer. No changes were recorded in average diurnal temperatures. However, extreme temperatures were affected. For temperatures below 0°C the regulation has caused an average drop of 0.7 ± 0.5°C, and for temperatures above 10°C there was rise due to regulation.

5. The annual temperature has been raised by some 0.1 - 0.2°C, mainly due to the reduced discharge of cold meltwater in spring and early summer.

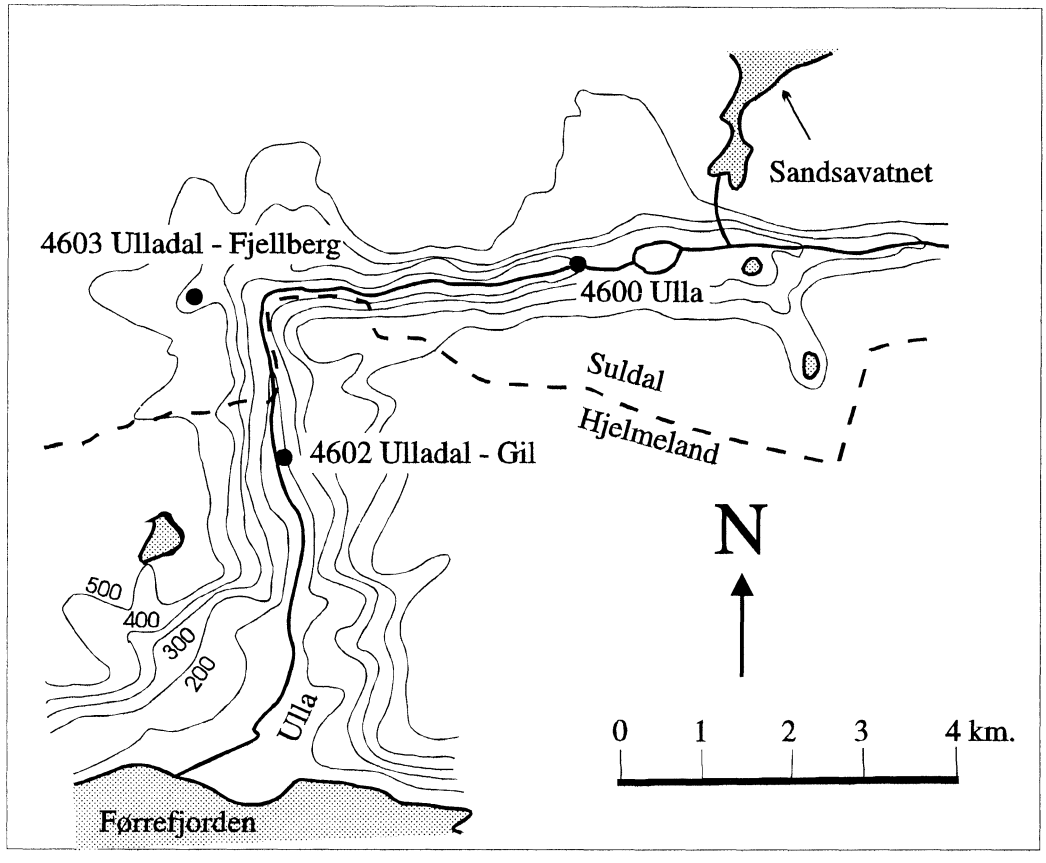


*Figure 3.17. Topographical map of the lower part of Ulladal Valley (after Nordli, 1988).*

Estimates based on general knowledge of micro- and local meteorology may also give a reasonable representation of expected changes.

The more rapid rise in water temperature in spring and the somewhat higher temperature in summer will influence ambient air temperature positively, especially where weirs (natural or artificial) form basins. The effect is strongest in cool, calm nights, e.g. when the air temperature drops below 0°C. Then, a reduction in water area has an opposite effect, and the frost alleviatory effect of a river can be diminished by a regulation. In daytime both higher water temperature and reduced water area affect ambient air temperature positively, but fairly large changes are required to induce significant effects.

In autumn (August-September) both the more rapid decrease in temperature and the reduction in water area may increase the incidence of frost on radiation nights, whereas daytime air temperature will rarely be affected. Complete drainage of a wide river bed in a terrain liable to radiation frost - i.e. fairly level, lowlying areas where cold air tends to settle on calm, clear nights - will induce the greatest changes. Nordli (1974 and 1975) calculated the frost protection effect of a river by means of a simple model developed by Høgåsen (1974). The transfer of sensible heat, $Q_H$, from the river to the air was given by:

$$Q_H = K(T_w - T_a) \text{ W m-2} \quad (\text{kcal m}^{-2} \text{ hour}^{-1}) \qquad (3.1)$$

where $K$ is a convection coefficient, $T_w$ is the temperature of the water surface, and $T_a$ is the air temperature. Conducting an experiment with very large temperature differences between water and air Høgåsen found $K$ values between 15 and 20 kcal $m^{-2}$ °$C^{-1}$ $hour^{-1}$, i.e. between 17.5 and 23.3 W $m^{-2}$ °$C^{-1}$. This may be the correct order of magnitude of $K$ at temperature differences of 50 °C or more. However, since $K$ increases markedly with increasing difference between warm water and cold air [cf. eqn. 4A in the Appendix, where $K$ is

proportional to $(T_w - T_a)^{1/3}$], the application of the values above for a temperature difference of, say 6°C would obviously give rather unrealistic values of heat transfer (cf. Figure 2.4).

Furthermore, it is assumed that the sensible heat transferred from the water - and possible added released latent heat due to condensation (sublimation) - is evenly distributed in the valley atmosphere below a selected convection level. On radiation nights there is nearly always a down-valley wind along a river. Therefore, due to the heat transfer from the water, the frost ameloriation is most pronounced above areas exposed to onshore airflow. But even complete drainage of Norwegian watercourses in question, would hardly lower the minimum temperature on a radiation night in autumn more than 1-1.5°C at the riverbank of the most exposed areas. At some 50-100 m distance from the bank no measurable effect would be likely (cf. conditions at Gransjøen, Chapt. 3.2.1). Therefore, results of such model calculations are not discussed here.

*Figure 3.18. Orkla at Nergard in Kvikne. Here the regulation would have drained most of the riverbed, but an artificial weir keeps the water level close to the natural state.*

Along comparatively flat stretches of the River Orkla upstream of Ulset power station in Kvikne (cf. map, Figure 3.23), several weirs have been constructed. Nevertheless, the water area is reduced some 30-50% by the regulation. Utaaker (1989) assessed consequential changes in ambient air temperature, and concluded that in spring and summer the positive effects of increased (higher) water temperatures are on the whole greater than the negative effects of reduced water area. On radiation nights in autumn (September) the minimum temperature may probably become some 0.5°C lower than prior to regulation at lowlying banks of the river, whereas at a distance of some 50 m from the water measurable changes will hardly occur. Along several parts of the riversides scrubs (Figure 3.18) shelter fields from influx of "river air", but the scrubs also retard the flow of cold air down the sides of the valley. Removal of those scrubs would undoubtedly raise the minimum temperature markedly over wide areas on radiation nights. Along steeper parts of the river, even near total depletion of the water flow has little or no effects on ambient air temperature during the growing season.

### *3.3.3. Consequences of changes in ice condtions*

In small, shallow lakes, weir basins, and slow flowing parts of a river, reduced waterflow may cause earlier freezing in autumn, more stable ice conditions in winter, and delayed ice break-up in spring.

Boe (1979) found that the construction of the Brokke Power Station has probably brought about an average delay of some 3 weeks in ice break-up on a fairly large, natural weir basin, Flåren (Figure 3.19), in the River Otra. Table 3.3 is based on temperature rcordings at Sandnes (8 m above the water level, some 70 m from the shoreline), and Bjørgum (5 m above the ice-free river, some 30 m from the bank). Three periods of 4 pentades ***before*** and 4 pentades ***after*** ice break-up on Flåren in the springs of 1976, 77, and 78 are included. In every spring the pentade when all ice was thawed, was discarded.

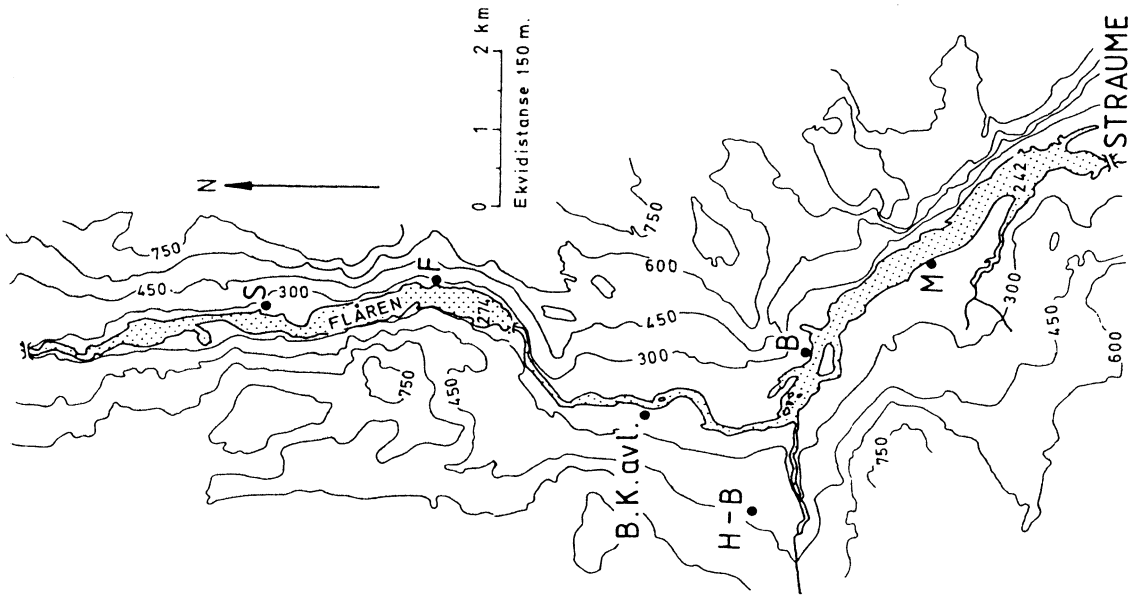


*Figure 3.19. Map of the area around Brokke Power Station (B.K) in the River Otra. Stations recording air temperature: Bjørgum (B), Sandnes (S), and vertical profile at Flåren (F) (Utaaker, 1982a).*

*Table 3.3. Mean differences, $\Delta_{A-B} = \Delta T_{after} - \Delta T_{before}$, in the mininimum temperature (°C) between Sandnes and Bjørgum by ice thaw. Data from 4 pentades* ***before (B)*** *and 4 pentades* ***after (A)*** *ice break-up on Flåren in the springs of 1976, 77, and 78. N is number of observations, ΔT is mean difference between Sandnes and Bjørgum, and $\Delta_{kr}$ = the lowest significant $\Delta_{A-B}$ at the levels : $^{+}$ = 95% , and $^{++}$ = 98% (Utaaker, 1982a).*

| | $N_B$ | $N_A$ | $\Delta_{A-B}$ | $\Delta_{kr}$ |
|---|---|---|---|---|
| Pentades | 12 | 12 | 0.77 | $0.76^{+}$ |
| Selected days | 26 | 31 | 0.96 | $0,91^{++}$ |

On the average for the pentades ice break-up caused a relative rise in the minimum temperature at Sandnes of 0.77°C - statistically significant at the 95% level (cf. Appendiks 6.1). For selected

radiation nights the average rise was 0.96°C - significant at the 98% level. Recordings of vertical temperature profiles at Flåren (Eitrheim, 1979) indicated that no effects due to changes in ice conditions were apparent at levels higher than 20-25 m above the surface of the basin. The more rapid rise in water temperature in Flåren subsequent to ice thaw after than before regulation, may also contribute slightly to the relative rise in ambient air temperature. The water temperature at Bjørgum is influenced strongly by that of the discharge from Brokke Power Station, where the rise in temperature in spring and early summer is rather slow.

Similar calculations for the maximum and the mean diurnal temperatures revealed no significant changes induced by ice thaw. It was concluded that the regulation had increased the incidence of frost in spring in lowlying areas at Flåren, but since ice break-up also after regulation, occurs normally before the start of the growing season, the growth-climate of the adjacent land is hardly affected measurably.

In the area at Lake Tveitevatn-Grungevatn (537 m a.s.l.) air temperatures were recorded at several heights above the water level during a period before and after ice thaw in the spring of the years 1971-74. The main results depicting the cooling effect of ice versus open water are summarized in Figure 3.20.

On radiation nights the cooling just above the ice surface (1.5 m) lowers the minimum air temperature some 2.0°C. The cooling effect decreases rapidly with elevation to some 1°C at 15 m, and some 0.3°C at 30 m above the lake. In some cases there was apparently an effect up to the level of 60 m, but significant lowering of the minimum temperature was not recorded at higher levels than 30-40 m above the frozen lake.

On calm, clear days the cooling effect of the ice lowered the diurnal mean temperature some 1.3°C at 1.5 m, 0.7°C at 15 m, and no effect appeared at 30 m hight above the ice surface.

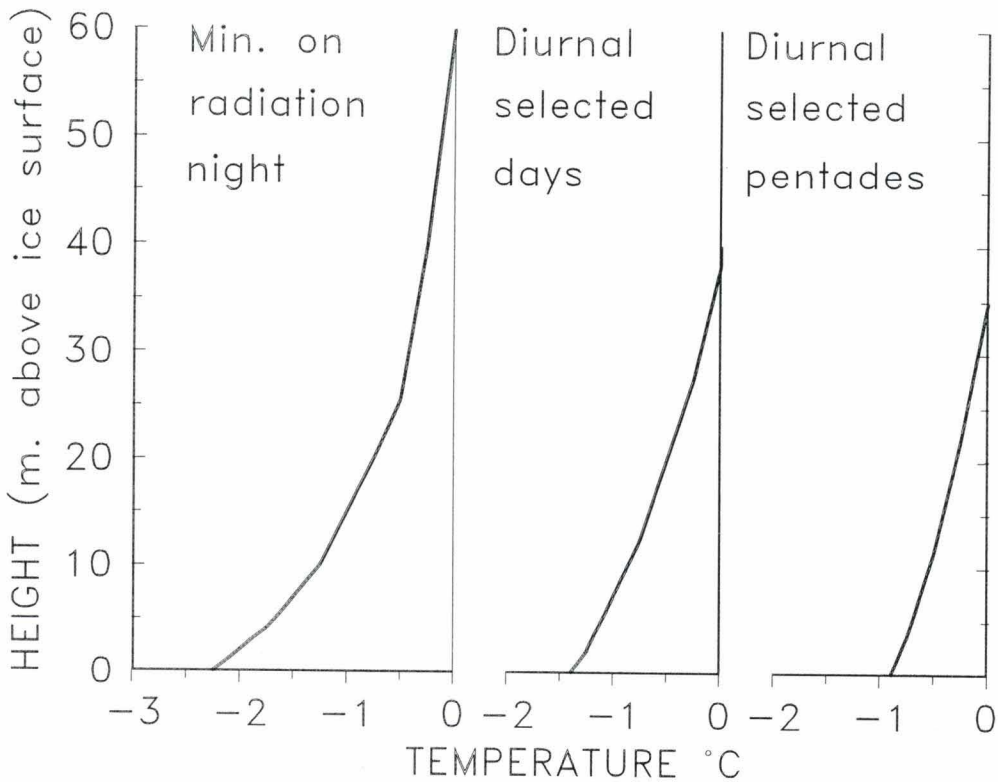


*Figure 3.20. Schematic representation of the cooling effect of ice-covered versus open water along the sides of the valley at Lake Grungevatn-Tveitevatn in Telemark by delayed ice break-up in spring (Utaaker, 1976).*

During extreme "fine weather" pentades before ice break-up an average lowering of diurnal mean temperature of some 0.7, 0.4, and 0.0°C can be expected at the levels 1.5, 15, and 25-30 m above the ice surface. According to Boe (1982) regulation has caused a lag of approximately one week in ice break-up at Lake Grungevatn-Tveitevatn in spring. Provided this is a "fine weather" week, this implies:

1. Increased incidence of frost - in a zone from the surface of the

lake and 30-40 upslope - on calm, clear nights when the air temperature drops below 0°C.

2. An average lowering of the diurnal mean temperature of some 0.5° in a belt up to some 30 m, and no measurable effect higher upslope.

In cloudy weather and/or by moderate or strong winds these effects disappear.

Experts assessing the effect of the regulation on plant growth estimated an average lag of 1-2 days in the growth of grass in a zone from the shore and some 30 m upslope.

At Lake Mjøsa, the largest lake in Norway, where ice conditions are not affected by regulation, Strandenes (1984) found that ice-covered versus open waterbody affects ambient air temperature strongly. He divided metorological data from 4 stations for the years 1965-74 in groups both according to temperature, and according to cloud cover and wind speed at a reference station. By means of the temperature difference method (cf. Appendix 6.1) he found that both classifications gave unambiguous results. Subsequent to ice break-up in spring the average relative rise in the minimum temperature on radiation nights was some 2.0°C at Kise (6 m above Lake Mjøsa, approximately 70 m from the shore), and some 1.5°C at Staur (31 m above the lake, approximately 600 m from the shore), whereas at Østre Toten (148 m above the lake) no effects appeared. For the average maximum temperature ($T_X$) the maximum rise after ice break-up was about 1°C at Kise and about 0.5°C at Staur - in the group $T_X < 10$°C.

These three examples indicate that the relative rise in the air temperature in the shore zone subsequent to ice break-up in spring is more pronounced around large lakes than at small lakes and weir basins.

### *3.3.4. Consequences of changes at waterfalls*

A waterfall usually generates a rather unique climate in its vicinity, and some distance downstream of the fall. The falling water entrains air, and spray is thrown up in the wind induced by the rushing water. The spray adds precipitation to an area in the wind direction. When the water is colder than the air, the air is cooled. Heating and evaporation of the waterdrops "draw" sensible heat from the air. When the water is warmer than the air, energy for evaporation is drawn from the water and sensible heat is also transfered to the air. On fine days during the growing season the environment influenced by the waterfall climate - in addition to spray precipitation - usually experiences lower daytime temperatures, higher humidities, and higher nighttime temperatures than the surroundings. If a waterfall is harnessed, the amount of spray is reduced strongly, or completely extinct, and the disappearance of the waterfall wind may increase the risk of frost at localities liable to stagnant air on calm, clear nights in spring and autumn.

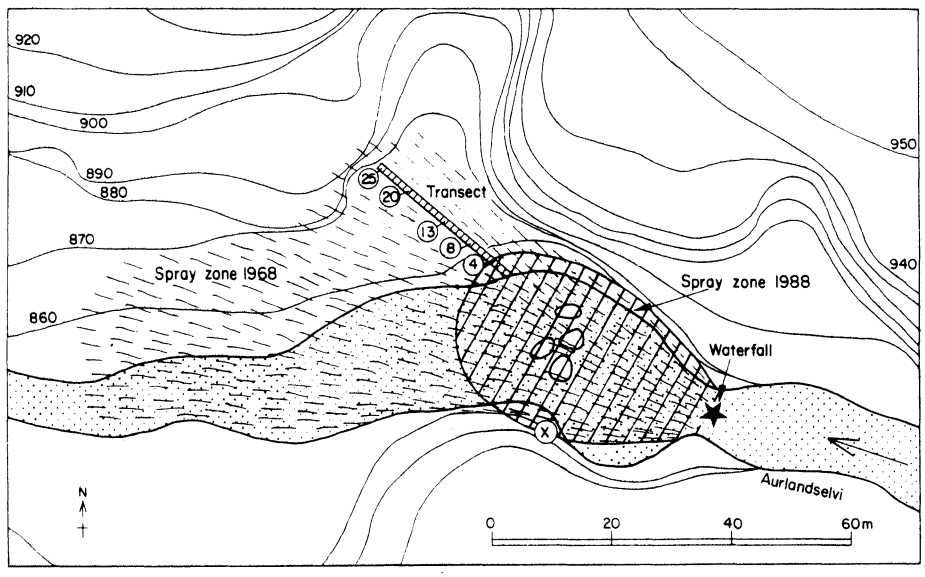


*Figure 3.21. The waterfall and a schematic representation of the investigated transect in Aurlandsdalen. The numbered circles are sites for recording of spray precipitation, and partly of air temperature (Odland et al., 1991).*

Odland (1990) and Odland et al. (1991) studied climate and vegetation at a waterfall (880 m a.s.l.) in the Aurland River before regulation, and again 20 years later. The water falls some 20 m directly into a pool. Brief series of measurements of spray precipitation and air temperature were made in a transect along the spray gradient of the waterfall (Figure 3.21). Based on these measurements and average rate of discharge, Odland et al. (op. cit.) estimated total monthly and annual spray precipitation. Before regulation (discharge 20-50 $m^3 s^{-1}$) the amounts for the months June-August (annual total) were 4870 (5160), 3520 (3640), 2440 (2530), 1510 (1530), and 230 mm (235 mm) at site 4, 8, 13, 20, and 25 respectively. After regulation, which reduced the discharge by some 92%, corresponding amounts were 90 (90), 15 (15) and 0 (0) mm at site 4, 8, and 13 respectively.

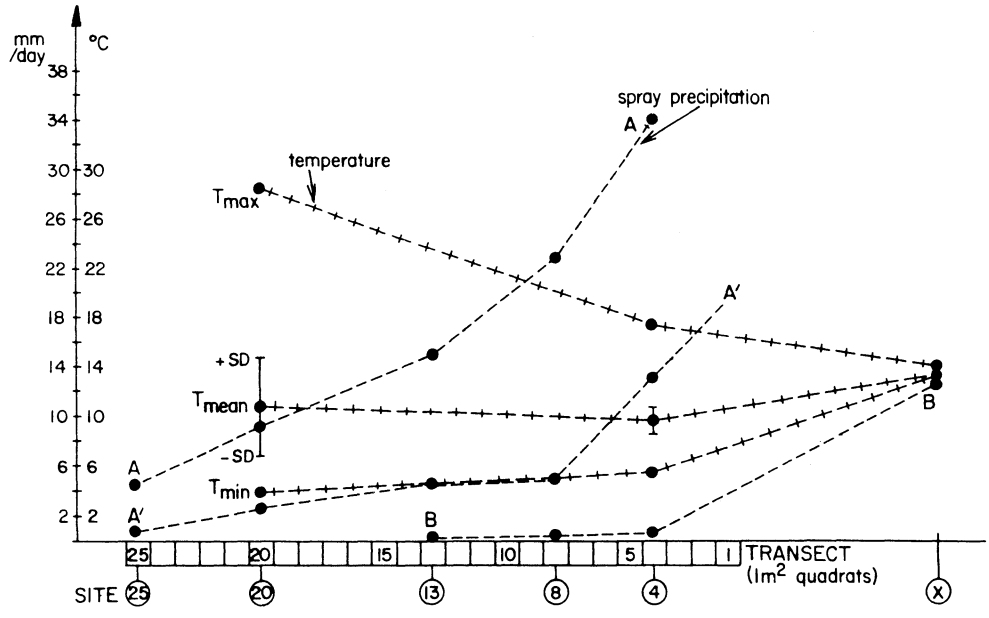


*Figure 3.22. Temperature and spray conditions along the transect (cf. Figure 3.21). Mean temperatures and standard deviations and extreme temperatures (recorded through 2-15 July 1988 at sites X, 4, 8, and 20) are shown. Spray precipitation measured at sites 4-20 in 1968: (A: 3-11 August, A': 27-30 August) and at sites X, 4, 8, and 13 in 1988 (B:4-8 July). The waterfall is approximately 20 m east of site X (Odland et al., 1991).*

Air temperature in the waterfall area was not measured before regulation. Recordings through 14 days in July 1988 documented significant temperature differences between sites (Figure 3.22). At site X, distanced some 20 m from the waterfall, where spray is still abundant, the temperature varied little between day and night and the relative humidity stayed mostly close to 100%. The effect of the waterfall on ambient air temperature is also evidenced clearly at site 4 where the amount of spray after regulation is small but significant.

During calm, cold weather in winter reduced discharge of a river reduces the deposition of hoarfrost, and the frequency and extent of frostsmoke in the vicinity of waterfalls and rapids that in the natural state were ice-free, while the main river was frozen up.

## *3.4. Rivers downstream of power stations*

As a rule the development of hydropower reduces the rate of discharge in the river downstream of a power station in summer and autumn. This especially applies to incidences of flooding due to rapid snowmelt and/or heavy rainfall in the catchment area of the reservoirs. In winter the discharge from power stations in general augments the waterflow markedly.

During the summer months June-August cold water discharged from a power station frequently causes considerably lower water temperatures than those prior to regulation. This effect is usually most pronounced in midsummer when a temperature drop of some 10°C has been recorded on warm days just downstream of power stations (Roen, 1988). As a rule the drop is far less, and local heating reduces the differences in water temperature between before and after regulation with distance downstream of the power station.

On warm summer days with light winds, the lower water temperature may occasionally result in somewhat lower air temperatures over

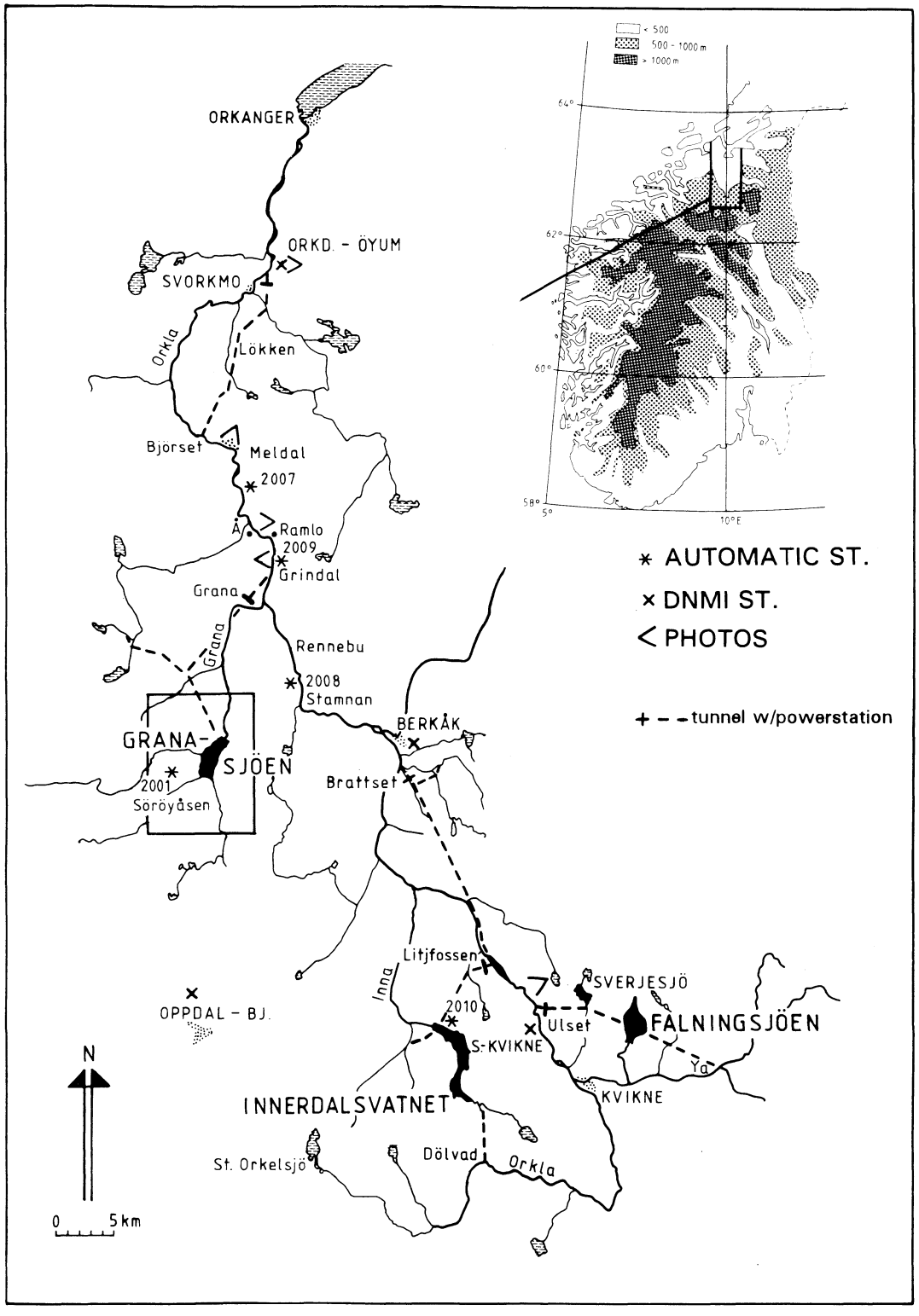


*Figure 3.23. Temporary, automatic, and permanent (DMNI) weather stations, tunnels with power stations, and positions for photographing frostsmoke in the Orkla/Grana region (Utaaker and Skaar, 1986).*

adjacent land down-wind of the river. However, in the surface layer the effect is usully small, because solar heating of the ground (and wind) generates intense convective mixing with the air above.

In late autumn - from about mid-September - the water discharged from a power station is usually warmer than the corresponding flow in the river before regulation. This may reduce the risk of frost along lowlying riversides, and also slightly enhance the formation of local fog.

In winter increased waterflow, and may be some 2-3°C higher water temperature than prior to regulation, will result in later, reduced, or no freezing of rivers and small, shallow lakes, downstream of power stations. Where the discharge from a power station poures directly into an ice-covered lake or fjord a lead always forms. Changes in ice conditions due to regulations are without doubt the major cause of changes in local climate along watercourses in Norway.

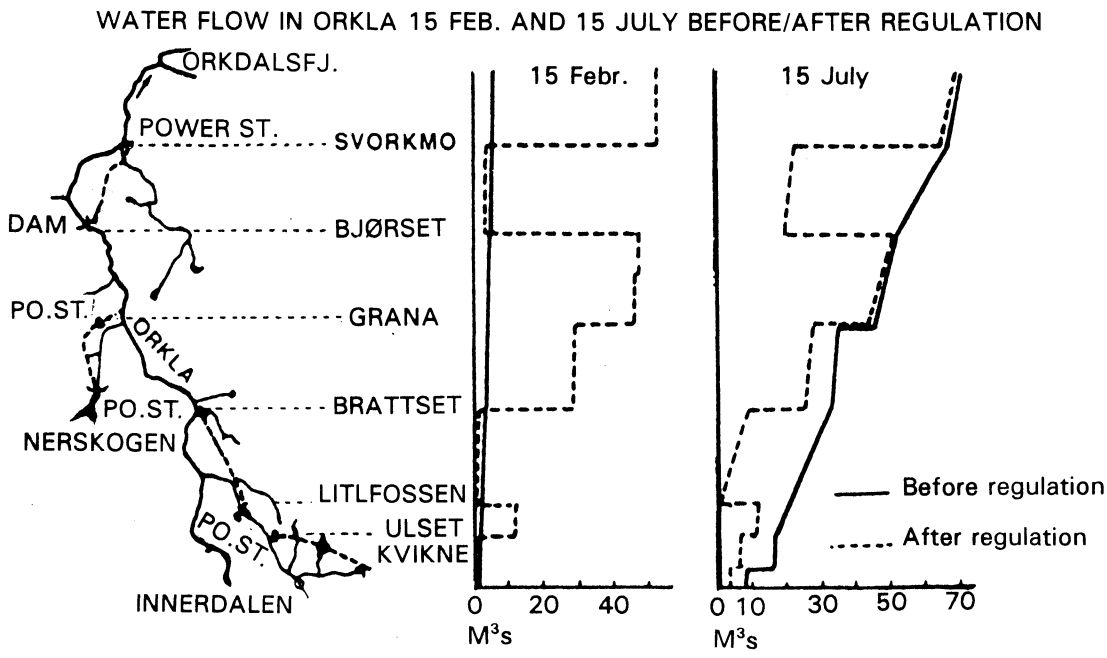


*Figure 3.24. Average waterflow in the River Orkla on 15 February and on 15 july* ***before*** *and* ***after*** *regulation (after Roen, 1988).*

In connection with the development of hydropower of the watercourse Orkla/Grana extensive investigations on the local climate in the Orkdal Valley were conducted both ***before***, and ***after*** the regulation was implemented. The schematic map (Figure 3.23) shows the location of climatic stations and power plants. It is seen that the automatic stations Stamnan (2008), Grindal (2009), Meldal (2007), and the weather station Orkdal-Øyum are located downstream of power plants. Figure 3.24 depicts average changes in the waterflow on 15 February and on 15 July due to regulation along different parts of the River Orkla.

### *3.4.1. Effects on water temperature and ice in the River Orkla*

Briefly summarized, Boe and Roen (Reports, 1982-1988) found that the regulations have induced marked changes in the water temperature downstream of the power stations during the growing season. The greatest changes occur during the summer months June-August, whereas the changes are small in April, May, and September. The estimated average lowering of the temperature when the power stations are in operation during the summer months are:

***Ulset Power Station-Litjfossen***: Drop at the outlet 2-3°C, and decreasing downstream.

***Brattset Power Station-Grana Power Station***: Drop at the outlet 1-2°C, and decreasing downstream.

***Grana Power Station-Bjørset***: Figures not stated, but it is said that when the station is in operation, the water temperature is considerably lower than before regulation.

***Downstream of Svorkmo Power Station***: Small changes, but a tendency to somewhat lower temperatures.

Assessing ice conditions along these parts of the river Boe and Roen (op. cit.) conclude (cf. Figure 3.20):

*Ulset Power Station-Litjfossen*: Prior to regulation, stable ice cover (except at rapids) all winters. Subsequent to regulation, no freezing at air temperatures above some -10°C. At lower temperatures some ice along the river banks espcially in the area towards Litjfossen.

*Brattset Power Station-Grana Power Station*: Freeze-up of this stretch of the river in its natural state normally started in November and was completed in December. Quite often spells of mild weather discontinued the freezing both once or twice in early winter, before a stable ice cover was established. Occasionally, extreme spells of mild weather could affect the ice conditions also later in the winter. The break-up of the ice occured usully in April. After regulation the ice cover of long stretches will normally break up several times during a winter, in step with spells of mild weather. When Brattset Power Station is in operation, the stretch some 2 km downstream of the outlet will not freeze even during extreme cold spells. In general, the entire river will have markedly more open water after regulation than before.

*Grana Power Station-Bjørset*: In the natural state conditions similar to these between Brattset and Grana power stations prevailed. When Grana Power Station is in operation, the river is always ice-free downstream towards Ramlo, whereas further downstream it freezes up and breaks up in step with fluctuations in the weather - mainly in air temperature. The time lag in the break-up of the ice during a mild spell is much shorter than before regulation. The break-up is most sudden and most complete in the upper part of this stretch, but also in the lower part conditions have changed markedly, except at the small reservoir (intake dam) at Bjørset.

*Downstream of Svorkmo Power Station:* Prior to regulation there was usually a stable ice cover throughout the months December-March, except at rapids where the ice around small leads could break up during mild spells. Subsequent to regulation most of

the river may still freeze up. However, due to the large increase in the discharge (cf. Figure 3.24) the process takes more time, and the leads at rapids will remain considerably wider.

### *3.4.2. Air temperature and humidity during the growing season*

As stated above, the regulation has caused a lowering of water temperatures downstream of the power stations in the River Orkla in summer. However, an analysis of recorded data by means of the difference method (Appendix, 6.1) revealed no measurable changes in ambient air temperature at any of the four meteorological stations in the Orkdal Valley (cf. Figure 3.23) due to changes in water temperature or waterflow. This result was not unexpected, since physical reasoning indicates that a drop of some few °C in the surface temperature of those comparatively small waterbodies should induce only small changes in ambient air temperature. However, if measurements had been made e.g. on the river bank immediately downstream of Grana power station on warm summer days prior and subsequent to regulation, a significant lowering in the air temperature had probably been found. Nevertheless, under these conditions convective mixing above the sun-heated ground erases the effect of the cool water surface on the air temperature rapidly as the air flows ashore. Possible changes in the micro-climate at low-lying areas along the river due to changes in the groundwater level were not assessed in this connection (cf. Chapter 3.3.1).

In summer evapotransporation from the vegetation is the main local moisture scource for the valley atmosphere. On warm days when the rate of evapotranspiration is high and the temperature of the river water is considerably lower than that of the air, evaporation from (condensation onto) the water surface is insignificant. A drop in the water temperature of some few °C has hardly a measurable effect on the process. On cool nights such a drop will reduce possible evaporation from the river slightly (cf. Table A2, Appendix).

As expected, no effects of the regulation on the humidity of the air were revealed along the River Orkla during the growing season. Nor at any other watercourse regulations in Norway have there been recorded measurable or visible (fog) effects on the local climate downstream of power stations due to changes in water temperatures and discharge during the growing season.

### *3.4.3. Air temperature and humidity in winter*

Changes in ambient climate are caused by the increase in the area of open (ice-free) water due regulation. Ice conditions in the River Orkla were discussed in section 3.4.1, and it was shown that in the river just downstream of a power station in operation, most of the water area is always ice-free. Further downstream parts that were previously normally frozen up throughout the winter months, now often break up and freeze in step with fluctuations in the air temperature. The area of open water which in general is large during mild or moderately cold spells, decreases rapidly during extremely cold spells.

Recorded temperature and humidity data were analysed by means of the difference method (Appendix, 6.1). The effect of more open water on ambient climate in winter is most evident in calm, clear weather, and it is expexted to increase with decreasing air temperature. The data were therefore grouped according to wind speed and temperature at the reference station, (1) Sørøyåsen (2001 in Figure 3.23, 800 m a.s.l.). In the example depicted in Figure 3.25 the groups are (T°C): I. $T > 0$ ; II. $-10 < T \leq 0$ ; III. $-15 < T \leq -10$ ; IV. $T \leq -15$ , and mean hourly wind speeds at Sørøyåsen and at (7) Muan (2007, 135 m a.s.l.) were below 1 m $s^{-1}$. At (8) Stamnan (2008, 213 m a.s.l.) and (9) Grindal (2009, 180 m a.s.l.) wind speeds frequently exceeded this threshold.

It is noted that, except in group II at station 9 after regulation, $\Delta T$ is systematically negative. Since the altitudes of the valley stations are from 587 to 665 m lower than that of the reference

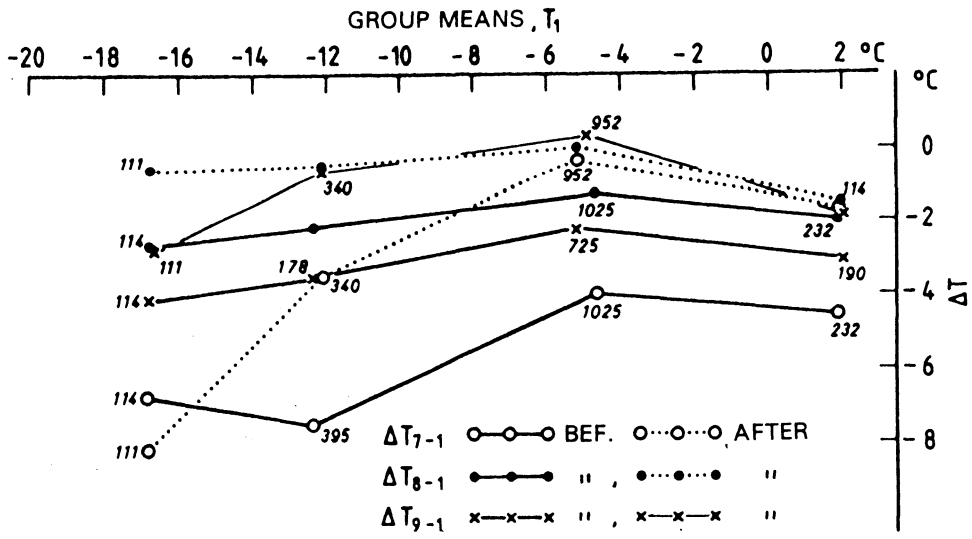


*Figure 3.25. Group means of the air temperature at Sørøyåsen, $T_1$ , and corresponding differenses, $\Delta T$ , referred to this station for Muan, 7 , Stamnan, 8 , and Grindal, 9 , during calm weather in January-February* ***before*** *and* ***after*** *the regulation of the watercourse. The number of observations in the respective groups is given (after Utaaker and Skaar, 1986).*

station, this indicates that irrespective of temperature level, a marked valley inversion prevailed in calm weather in winter both before and after regulation. The regulation has evidently caused a rise in air temperature at the three valley stations. At Stamnan (8) the temperature rise in group I is, as expected, small and statistically insignificant, whereas in the groups with lower temperatures it is statistically significant, and increases from 1.26°C in group II to 2.07°C in group IV. This result agrees with the effect Utaaker (1982a) found at the River Otra downstream of Brokke power station.

The large apparent rise in air temperature at Grindal (9) and Muan (7) in group I, due to more open water in the river, cannot be

explained physically. Here the relative temperature rise in group II is also considerably higher than the effect of ice versus open water that was recorded in the corresponding temperature group by Pleym (1980) at Lake Storsjøen in Rendal, and by Strandenes (1984) at Lake Mjøsa. The recorded effect in group III at stations 7 and 9 is in better agreement with the results from these investigations, where a relative temperature rise of 4.9°C or more were found for temperatures below -10 and -15°C respectively. However, it should be noted that the areas of those waterbodies are much larger than the areas in question in the River Orkla.

Smaller effect at station 9 in group IV compared to group III is probably due to partial or total freezing-up of slow-flowing parts of the river which remain ice-free at somewhat higher freezing temperatures. Then, reduced heat transfer from the river results in a smaller rise in ambient temperature. This effect is evident also at station 7, but it does not explain the apparent relative drop in the mean temperature in group IV subsequent to regulation.

Utaaker (1983) found that by application of the difference method, choice of reference station may have decisive effect on the results. A detailed study of some selected cases of calm or light winds, and rather low temperatures showed that the valley stations at Orkla were always much colder than that at Sørøyåsen. The strongest inversions developed at the lowest lying station, Muan (7), where the temperature of the stagnant cold air at times were more than 20°C lower than at the reference station Sørøyåsen. At Stamnan a moderate (2-4 m $s^{-1}$) mountain wind prevailed in these situations, and the air temperature was seldom more than some 5°C lower than that at Sørøyåsen. At Grindal both wind speeds and air temperatures were as a rule somewhat lower than at Stamnan, but markedly higher than those at Muan. This study showed that Sørøyåsen was a fairly representative reference station for Stamnan, less so for Grindal and rather unsuitable for Muan.

Nevertheless, based on the calculations with Sørøyåsen as a reference station, and reciprocal comparisons between the valley sta-

tions it was possible to assess the effect of the regulations on the air temperature at the three valley stations in winter with reasonable accuracy. Analogous analyses were carried through for Orkdal-Øyum (21 m a.s.l.), and for S-Kvikne (550 m a.s.l. , some 800 m upstream of Ulset Power Station). Conditions downstream of Ulset Power Station were estimated on basis of the results from the stations along Orkla, and from studies downstream of Savalen Power Station in Alvdal (Utaaker, 1978), and Brokke Power Station in Setesdal. Downstream Brokke Utaaker (1982a) found an average rise in ambient air temperature of some 2°C due to regulation at temperatures below -10°C.

Briefly summarized, one can expect the following effects of the Orkla/Grana regulation on ambient air temperatur by calm or light winds in winter when the power stations are in operation:

1. At temperatures about or above 0°C; no significant effects at any of the stations.

2. At temperatures about -10°C; an average temperature rise of some 1.3°C Stamnan (2008) and at Orkdal-Øyum, and some 1.5°C at Grindal (2009), at Muan (nr. 2007), and just downstream of Ulset Power Station.

3. At temperatures about -15°C; an average temperature rise of some 1.7°C at Stamnan and at Orkdal-Øyum, and about 2.8°C at Muan, Grindal and Ulset.

4. At temperatures about -20°C; an average temperature rise of some 2°C at Muan, Stamnan, Grindal, and Orkdal-Øyum, and some 3.5°C at Ulset.

5. At temperatures about -25°C or lower; a temperature rise of less than 1°C at Grindal and Orkdal-Øyum, and probably some 4°C at Ulset, and no effect at Muan. At Stamnan air temperature never dropped much below -20°C. (The reason why the effect of the regulation decreases at extremely low tempera-

tures is partial icing over of the area of ice-free water).

6. At S-Kvikne there were no measurable effects of the regulation on air temperature.

The recordings of the humidity of the air were analysed in a similar way. Nor for this element was Sørøyåsen an ideal reference station for the stations in the Orkdal Valley. The results indicated that regulation has caused no significant changes in relative humidity of the valley atmosphere. However, in calm, cold weather - when the rate of evaporation from open water is fairly large (cf. Appendix, Table A2) - a significant increase in absolute humidity (mass of water vapour per volume unit of air) was evidenced at the stations Stamnan, Grindal, Muan, and Orkdal-Øyum. However, due to the simultaneous rise in air temperature (documented above), and the fact that possible surplus of water vapour in relation to an ice surface (cf. Appendix, Table A1) is deposited as hoarfrost before the air enters the instrument screen, the recorded relative humidity of the air increases only slightly. Under these conditions outdoor air that e.g. ventilates a cowhouse, has as a rule somewhat increased "drying power" (saturation vapour pressure deficit) due to regulation. By very low freezing temperatures the changes both in absolute humidity and "drying power" of the ventilatory air is of no practical significance. Air humidity conditions at S-Kvikne are apparently not changed by regulation.

### *3.4.4. Frostsmoke*

The most striking and visible changes in local climate induced by development of hydropower in Norway are increased frequency and extent of frostsmoke, and augmented deposition of hoarfrost downstream of outlets from power stations. These phenomena are most fully developed in regions with frequent spells of calm, cold weather in winter, especially along parts of the watercourse where the regulation has impeded the natural freeze-up of the waterflow even by extremely low temperatures.

Systematic studies of onset and development of frostsmoke have been conducted at several Norwegian watercourses. The most comprehensive observations covering all days throughout 14 winter seasons (November-March) were made at the outlet of Lake Vågåvatn (Nordli, 1988). Based on these data and observations from a nearby weather station Nordli (op. cit.) developed several criterions for onset and development of frostsmoke. He found that a model based on a combination of air temperature and cloudiness yielded the closest correlation to the observations. It was, however, only slightly better than a model based on the *temperature difference between the water surface and the air* (cf. Chapter 2.4, Figure 2.10). By means of the latter model Nordli calculated longtime (30 year), average probabilities of frostsmoke at different hours of the day during these winter months. He found the highest probability at 0700 in the morning and the lowest at 1500 in the afternoon (Table 3.4).

*Table 3.4. The probability of frostsmoke at the outlet of Lake Vågåvatn during the months November-March, averages 1931-1960. $F_1$ = all frostsmoke that may be observed, $F_2$ = all well developed frostsmoke, $F_3$ = all frostsmoke that at least part of the time spreads beyond open water (after Nordli, 1988).*

| Diurnal variation | Time (h) | Cases per season $F_1$ | $F_2$ | $F_3$ |
|---|---|---|---|---|
| Most frostsmoke | 0700 | 44 | 33 | 17 |
| Least " | 1500 | 32 | 24 | 13 |
| Diurnal mean | 0000-2300 | 38 | 29 | 15 |

The regional climate is rather continental. Winter temperatures frequently drop below -20°C, and temperatures close to -40°C may occur. During cold spells wind speeds in the area is rarely above 2 m $s^{-1}$, and Nordli did not study the effect of wind speed on onset and development of frostsmoke. Observations downstream of hydropower stations in other Norwegian valleys (Utaaker, 1979) indicated that the probability of extensive frostsmoke decreases with increasing wind speed (cf. Chapter 2.4, Figure 2.6). At speeds above 2 m $s^{-1}$ frostsmoke rarely spreads markedly beyond open water.

*Table 3.5. The probability of frostsmoke downstream of Rendal Hydropower Station during the months November-March, averages 1965-74. Frostsmoke classes as in Table 3.4 (after Utaaker, 1980).*

| Time of day | Hour | Cases per season $F_1$ | $F_2$ | $F_3$ |
|---|---|---|---|---|
| Morning | 0700 | 30-40 | 27-32 | 8-10 |
| Noon | 1300 | 15-20 | 12-15 | 3-4 |
| Evening | 1900 | 20-30 | 17-22 | 6-8 |

Table 3.5 gives calculated probabilities of frostsmoke at the outlet canal from Rendal Power Station in Øvre Rendal Valley (Utaaker, 1980). Here the effect of the wind speed which frequently exceeds 2 m $s^{-1}$ during cold spells, is assessed.

The normal discharge from Rendal Power Station (cf. Figure 3.28) is some 40-50 $m^3$ $s^{-1}$, and the 40 m wide and some 3 km long outlet canal is always ice-free when the plant is in operation. Lake Lomnessjøen is normally frozen up from mid-November through April. The discharge from Lake Vågåvatn was mostly 40-60 $m^3$ $s^{-1}$, and the some 175 m wide River Otta was ice-free over a downstream distance of some 4 km. When Lake Vågåvatn was iced over, there was always a lead of some 200x500 m at the outlet. During the 14 seasons the earliest freeze-up ocurred on 14 November and the latest one on 1 January. Considerably larger area of open water at the outlet of Lake Vågåvatn than at Rendal Power Station under typical "frost smoke weather" is probably one reason why the probabilities of frost smoke - especially of $F_3$ - are markedly greater at the former locality. A somewhat colder winter climate in Rendal than in Vågå should have an oposite effect. Therefore, the main reason for those differences is most certainly that in calm, cold weather a stronger mountain wind develops in Øvre Rendal than along Lake Vågåvatn.

Freezing of rivers in the interior of Southern Norway in their natural state normally begins in November, and from early December through March-April a stable ice cover usually persists, except at

rapids and waterfalls. To assess changes in frostsmoke conditions in November due to regulation is therefore rather difficult. Extensive frostsmoke at regulated watercourses may occur in March at night and early in the morning, but it normally dissipates rapidly when the sun rises. Studies at several power plants have therefore been limited to the main winther months December-February. Provided the river stretch of current interest was frozen up in its natural state, all changes in frostsmoke conditions are then caused by the regulation.

In Table 3.6 are compiled results downstream of the power plants: Rendal (Utaaker,1980), Savalen in Alvdal (Utaaker, 1979), Trollheim in Surnadal (Utaaker, 1984), and Ulset, Brattset, Grana, and Svorkmo along the River Orkla (Utaaker and Skaar, 1986). Patches of "smoke" above open water or "smoke" that dissipates immediately above the water hardly give any trouble or annoyance to people living in the area. Moreover, these phenomena are frequently difficult to observe. Such cases are ommitted here. Thus the table gives frequencies of comparatively dense frostsmoke that can be recognized from points at some distance from open water. Average discharge from the power plants, or increase in the waterflow due to regulation is also given. The results are based on observations and calculations. It is noted that prior to regulation extensive frostsmoke occurred rarely during these winter months at any of those localities

At most localities the frequency of dense and extensive frostsmoke, $F_3$ , was greatest in extremely cold winters when air temperature dropped towards or below -20°C frequently. However, in winters with frequent, abrupt changes from mild, cloudy to calm, clear, cold weather, the incidence of $F_3$ may compare with that in extremely cold winters. This is due to the fact that by a sudden cooling of fairly still, moist air, water vapour is condensed into droplets forming radiation fog. Near open water this adds to developing frostsmoke and may result in a dense extensive fog blanket in the valley bottom. Moreover, the area of ice-free river surface is usually relatively large after a mild spell making the

*Table 3.6. Probable frequencies of frostsmoke during the months December-February - on the average, and in a winter with exceptionally high frequency - downstream of the power stations (pst.) Rendal in Øvre Rendal, Savalen in Alvdal, and Trollheim in Surnadal, and along the following stretches of the Orkla River:* ***Ulset*** *pst.-****Storfossdam****,* ***Brattset*** *pst.-****Grana*** *pst.,* ***Grana*** *pst.-* ***Grindal****,* ***Ramlo-*** ***Bjørset****, and downstream of* ***Svorkmo*** *pst. The increase i waterflow, Wf , downwards the river Orkla is mainly discharge from the power stations.*

| Power st./ stretch | Wf. $m^3s^{-1}$ | Average $F_2$ | Average $F_3$ | Exceptionally high $F_2$ | Exceptionally high $F_3$ |
|---|---|---|---|---|---|
| Rendal | 40-50 | 20-24 | 6-8 | 30-35 | 10-14 |
| Savalen | 20 | 10-13 | 2-3 | 20-26 | 5-6 |
| Trollheim | 27-30 | 9-14 | 6-9 | 17-22 | 10-14 |
| Uls.-Sto. | 12 | 8-12 | 2-4* | 17-25* | 8-12* |
| Bra.-Gra. | 30 | 3-5* | 1-2* | 6-9 | 2-3 |
| Gra.-Gri. | 50 | 7-11 | 2-4 | 16-22* | 6-8* |
| Ram.-Bjø. | 50 | 8-12 | 3-5 | 17-24 | 7-10 |
| Svorkmo | 53 | 7-11 | 2-4 | 14-19 | 5-7 |

** Corrections relative to Utaaker and Skaar (1986)*

supply of moisture to the air more abundant. Therefore, the most extensive and stable frostsmoke often develops under such circumstances. In calm, clear weather a fairly stable fog may develop above the open river and adjacent land (Figures 3.26 and 3.27).

By continuing cold weather the humidity of the cold air flow is low, the area of ice-free water is often gradually reduced on parts of the river, and in many valleys the mountain wind is well developed. Under these cirumstances intensive "smoke" from open water tends to dissipate both in low height above, and at a very short distance beyond the edge of open water (cf. hoarfrost pictures, Figures 3.29 and 3.30).

By calm weather in winter, air temperatures at the outlets of the power stations Rendal, Savalen, and Ulset are usually much the same. The higher frequencies of well developed frostsmoke at Rendal

Power Station, documented in Table 3.6, are probably mainly due to the greater discharge from that plant. However, more calm or light winds here than in Alvdal (Savalen) and Kvikne (Ulset) in these situations, may be part of the explanation. The fact that the incidence of $F_3$ is greater at Ulset than at Savalen in severe winters, may partly be attributed to somewhat stronger mountain winds in Alvdal than in Kvikne. However, unequal operation of the two power stations - frequent, discontinued running during nighttime at Savalen, which resulted in reduced area of ice-free water, and more stable operation at Ulset - may have contributed to these differences.

*Figure 3.26. Dense frostsmoke at Glomma as seen from Nordre Glomstad towards Åsta Bridge. The temperature difference between water and air was 22°C, and the wind speed some 0.7 m $s^{-1}$. The smooth top surface of the layer of fog indicates that intense radiative cooling has created an inversion "lid" that prevents the frostsmoke from rising to a higher altitude.*

*Figure 3.27. Frostsmoke over Orkla River downstream of Ulset Power Station, Kvikne 07.01.86. The temperature difference between water and air was ca. 23°C, and wind speed was ca. 0.5 m $s^{-1}$.* ***Below:*** *From Storeng Bridge at the outlet of the power station.* ***Above:*** *2-3 km downstream of the outlet (photos KVO).*

Comparatively well developed mountain wind, and relatively small areas of ice-free water are the causes of the low frequencies of frostsmoke on the stretch between Brattset and Grana Power Station. It is noted that the numbers are hardly representative of conditions for the uninhabited area from Brattset and some 2-3 km downstream, where no observations were made.

Also immediately downstream of Grana Power Station (Table 3.6, Gra.-Gri.) marked mountain wind curbs the development of frostsmoke frequently. In comparison Trollheim in Surnadal with much the same air temperatures, but lower wind speeds than Grindal, in calm, cold weather in winter, has considerably higher frequencies of extensive frostsmoke. For example, during the winter 1983/84 Grindal had more days with temperatures favoring development of intense frostsmoke than Surnadal. However, the number of days with observed $F_2$ and $F_3$ downstream of the power stations were 5 and 3 in Grindal, and 7 and 12 in Surnadal.

Between Ramlo and Bjørset much of the river freezes up during extremely cold spells, and then the development of frostsmoke decreases gradually. Simultaneously accumulation of cold air retards the flow, and nearly calm winds and considerably lower air temperatures than at Grindal prevail, causing more frequent, extensive, and persistent $F_3$ in this area than in Grindal. It should be noted that even by nearly still air, the frostsmoke is rarely stationary in the valley bottom. It forms continually above open water and dissipates at its edges, and sometimes the blanket of fog oscillates more or less fortuitously (cf. Figure 3.9).

The recordings of relative humidity of the air in the Orkdal Valley evidenced that most of the surplus of water vapour in the air (cf. Chapter 2.5), due to evaporation from the the ice-free river, was depleted before the air passed the hygrometer in the instrument screen. Obviously, it was deposited as hoarfrost on the ground or on obstacles in the airflow, and/or mixed into drier ambient air. This also implies that that all fog droplets were evaporated before the air entered the screen.

### *3.4.5. Hoarfrost*

Deposition of hoarfrost onto the ground and onto exposed objects, which is a common phenomenon also along ice-covered (nonregulated) watercourses (cf. Figure 2.11), is augmented when the area of ice-free water of rivers or lakes is enlarged by a regulation.

In Rendal Valley systematic measurements of hoarfrost deposition (on 5 square surfaces, 0.23 $m^2$ , 1 horizontal, and the others vertical exposed towards N, E , S , and W respectively) were made at 4 stations (Elvål, Løsåmoen, Kvernnesodden, and Åkrestrømmen, Figure 3.28) during two winters before and three winters after the regulation (Nybø, 1984). With Elvål (some 14 km upstream of the outlet of the power plant) used as a non-influenced reference station Nybø found the following, average effects of the regulation for selected days with minimum temperatures below -4°C, no precipitation, and light or no winds:

At ***Løsåmoen*** (some 3 km downstream of the power station and some 60 m from the 40 m wide canalized river, that is ice-free in winter whenever the turbines are running) a statistically significant increase in hoarfrost deposit of approximately 23 g per day over the sampling area (1.15 $m^2$), corresponding to 0.02 mm of water. For "hoarfrost days" with minimum temperatures below -15°C Nybø found an average increase of 25.4 g. His findings corroborate the result presented in Table 3.7 which is based on a sample of the same data (Utaaker, 1982).

At ***Kvernnesodden*** (some 7 km downstream of the ice-free canal, at the southern shore of Lake Lomnessjøen, where the regulation has caused only negligible changes in the ice conditions) no increase in hoarfrost deposit was evidenced.

At ***Åkrestrømmen*** (some 350 m from Rena River which here flows ice-free when the turbines are running, and some 600 m from Lake Storsjøen where a lead prevail when the lake is iced over) the average diurnal increase was some 19 g, i.e. some-

what smaller than that at Løsåmoen, and the significance level was 95%.

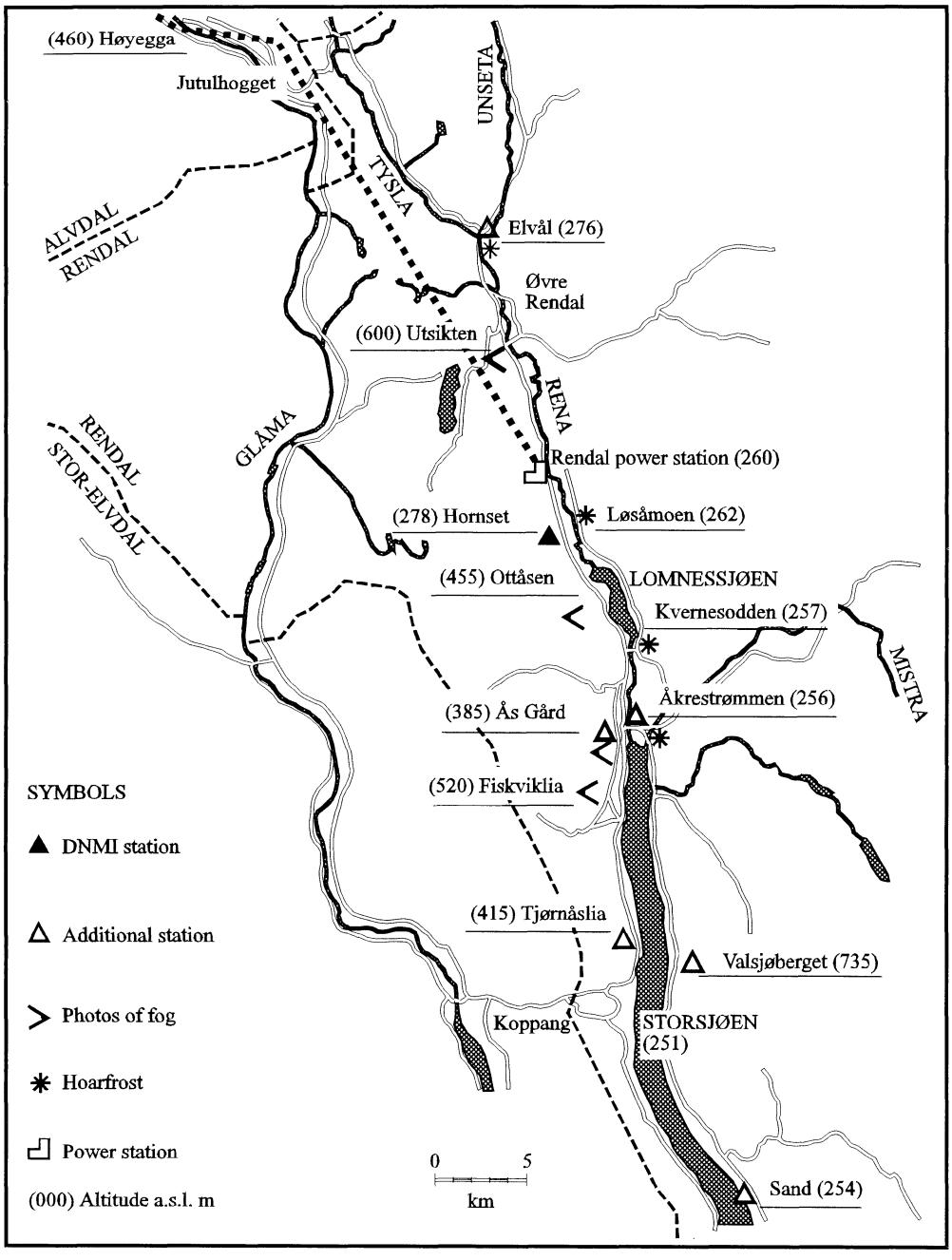


*Figure 3.28. Map of stations in Rendal Valley (after Pleym, 1980).*

*Table 3.7. The effect of the regulation on average deposition of hoarfrost, (g per 1.15 m²) at Løsåmoen (L) downstream of Rendal Power Station in relation to Elvål (E), in calm weather with air temperatures below -10°C. N = number of days, $\Delta_{L-E}$ = mean difference, $\sigma^2$ = variance, and $^{xxx}$ = significans level 99.9% (after Utaaker, 1982b).*

| | N | E (g) | $\Delta_{L-E}$ (g) | $\sigma^2$ | L/E % |
|---|---|---|---|---|---|
| Before reg. | 35 | 35.0 | 12.9 | 726 | 127 |
| After " | 49 | 20.6 | 36.7 | 613 | 309 |
| $\Delta_A - \Delta_B$ | | | $23.8^{xxx}$ | | |

This indicates that the accretion of hoarfrost decreases with the distance from open water. However, especially at localities where an onshore flow of moist air invades the area, the visible effect may at times be observed at a distance of some 2-3 km from the ice-free river (lake). The documented increase near open water was most pronounced by low temperatures when well developed frostsmoke invaded the area. But also in situations with no visible frostsmoke a significant increase in hoarfrost deposit due to the regulation was documented. At a water temperature of e.g. 1°C the evaporation from open water will contribute noticeably to the vapour content of the air when the air temperature drops below some -5°C (cf. Table A2, Appendix). By "hoarfrost weather", i.e. calm, clear weather, a statistically significant increase in hoarfrost deposit due to adjacent, ice-free water can be expected whenever subfreezing air temperatures drop some 5°C or more below the temperature of the water.

Sharp edges and pointed and thin objects are more effective hoar-frost collectors than plane surfaces. Therefore, one must expect considerably larger amounts of hoarfrost on e.g. coniferous forest near open water than those found at the measuring stations in Rendal. Figure 3.29 depicts an example of extreme deposit on a wire fence and on forest close to the open river. Notice, that onto the vertical iron bar supporting the fence the frost is typically accreted mainly on the "windward" side, i.e. against the gentle air

flow along the river. The frostsmoke picture from Storeng Bridge (Figure 3.27) also indicates vigorous frost deposition onto the forest along the river brinks.

It is worth mentioning that the largest frost deposit at Løsåmoen - 182 g (water equivalent 0.16 mm) - was measured after a day (24 hours) with $T_N$ = -3.3°C. Corresponding values at Åkrestrømmen were 184 g, and $T_N$ = -4.2°C. These extreme values coincided with a sudden change from mild humid, to clear, calm, cold weather. The deposition of frost was then mainly due to moist, maritime air which had been advected into the region.

*Figure 3.29. Hoarfrost and frostsmoke at Åsta Bridge on Glomma River as seen towards north, 07.12.81. The temperature difference between water and air was ca. 22°C and there was a weak flow from north. Visible clouds are stratus developed above the ice-free Lake Storsjøen. Towards south the sky was clear (photo A. Boe).*

*Figure 3.30. A birch tree covered by hoarfrost, and frostsmoke above the Orkla River at Muan 20.01.85 (photo E. Skaar).*

It is noted that the most vigorous deposition of frost usually occurs by sudden changes from mild to cold weather, and vice versa. By sudden changes to mild weather the surface temperature of the ground and other objects usually rise markedly slower than the temperature of the air. As long as the surface temperature is below 0°C and the dewpoint temperature of the air is higher, hoarfrost is deposited.

A case of heavy frost on a birch tree is depicted in Figure 3.30. This is easily explained by preceding weather. On 18 January the air temperature at Muan reached 1.8°C. Simultaneous relative humidity was 97%, and a moderate wind prevailed. During the next day abating wind and clear skies caused a considerable temperature drop (-10.5°C at 0800 h, and -23.3°C at 2300 h), and extensive fog formed in the valley. The combination of radiation fog and frost-

smoke persisted during the calm (wind speed below 0.5 m $s^{-1}$), cold weather untill about 1200 h on 20 January. Then a light wind caused a minor rise in the air temperature and most of the fog dissipated, restricting the frostsmoke to the area of ice-free water.

Most of the frost on the birch was sublimated from the vapour in the mild air, and from the fog droplets which also mainly originated from the vapour advected from the ocean. The contribution by local evaporation from the ice-free river was probably of minor importance.

Measurements in Hallingdal Valley (Kanavin, 1968) indicated an increase in frost deposit downstream of Hol Power Plant well in accordance with the amounts recorded in Rendal.

During a typical hoarfrost day with air temperature -20°C, relative humidity 70%, and wind speed 0.5 m $s^{-1}$ the evaporation from water with surface temperature 1°C is about 2 mm (Table A2, Appendix). Provided this water vapour is deposited as hoarfrost uniformly with 25 g $m^{-2}$ (water equivalent 0.025 mm), a local area of 80 times the area of open water would be affected. It should be noted that frost deposition does not start before the air is saturated in relation to an ice surface, which in this example implies a relative humidity of 82% (Table A2, Appendix). Moreover, even at wind speeds as low as 0.5 m $s^{-1}$ entrainment of drier air from above and from the sides of the valley will reduce the fraction of vapour that is deposited as frost in the environment of open water.

Measurable changes in frost deposit due to a regulation will accordingly be restricted to areas along the watercourse where augmented discharge in winter has enlarged the area of ice-free water. Since evaporation increases with increasing temperature difference between water and air, the increase in hoarfrost deposit in the surroundings of a water surface of constant area is largest in calm and extremely cold weather, when also dense frostsmoke normally forms.

## *3.5. Fjords receiving discharge from regulated watercourses*

When hydropower is developed in the catchment area of a fjord the amount of fresh water running into the fjord in winter usually increases. A change in ice conditions induced by this increase is probably the sole, conceivable effect on ambient climate.

### *3.5.1. On freezing of fjords*

The freezing of a fjord depends both on weather conditions and the stability of the surface layer of the water. The density of salt water increases with increasing salinity and decreasing temperature. In some cases a surface layer of low salinity forms. This layer of brackish water represents a stable stratification, and the stability increases with increasing difference in salinity between the surface layer and the saline water beneath. The brackish water originates from local precipitation and fresh water discharged from the catchment area.

For freezing to occur the water surface must be cooled to the freezing point. When a stable brackish surface layer cools off in calm weather, it remains lighter than the saline water beneath. The thermal convection is than restricted to the brackish layer, and a comparatively small heat loss may result in beginning ice formation. The surface is cooled by transfer of sensible and latent heat to colder air, and by long-wave radiative loss (cf. Figure 2.4). The radiative loss is most pronounced on clear nights (no gain of short-wave solar radiation). Wind creates waves and turbulens which disrupt the stable stratification of the surface layer, and thus prevents freezing. Calm winds and precipitation - especially snow that also cools a thin surface layer - followed by clear skies and no wind are ideal weather conditions for ice formation on a fjord. Under these circumstances the fjord may freeze up in one night even when air temperature measured 2 m above the ground on the shore remains above 0°C throughout the night. In winter - especially

when surrounding land is snow-covered - cold air drainage down the sides of the fjord and down adjacent valleys sets up a breeze that in many cases prevents ice formation.

When regulation increases the discharge of fresh water to a fjord in winter, this may strengthen the brackish surface layer, and enhance the stability which would develop naturally under pre-regulation conditions. Here it is presumed that the discharge from the power station pours into the fjord directly, or via a river. According to Boe and Roen (1991) "the continuous supply of fresh water makes it more difficult to disrupt the layer of brackish water by turbulent mixing, and increases the probability of hydrophysical conditions that make the surface liable to icing when the ideal combination of air temperature and wind occurs. Continuous inflow of fresh water also facilitates further growth of an established sheet of ice. In other words, a watercourse regulation entails a heigher risk of icing of a fjord, and frequenly also a more extensive, stable, and persistent sheet of ice."

In the Lustrafjord in Sogn (Figure 3.31) several studies have been conducted on ice conditions and the effects of the regulation of the watercourses Fortun/Granfasta (in operation from 1959), Leirdøla (from 1978), and Jostedøla (from 1989) on freezing and extension of the sheet of ice (Gjessing 1968, 1982, and 1985, Roen 1983, Boe and Roen, 1991). Different elements of the local climate in Luster are discussed by Gjessing (1968), Skaar (1976), and Utaaker (1979c). Comprehensive studies of the effects of ice-covered versus open water on air temperature and humidity were implemented by Hammer (1986), and by Gjessing and Nordli (1991).

### *3.5.2. Effects of regulations on ice conditions in the Lustrafjord*

Freezing of the fjord in its natural state (Gjessing, 1968, Boe and Roen, 1991) obviously depended on - mostly rather heavy - precipi-

tation and/or large discharge from rivers pouring into the fjord prior to favourable weather conditions. A combination of precipitation and enhanced feeding from rivers and brooks apparently created the ideal hydrophysical conditions for extensive freeze-up.

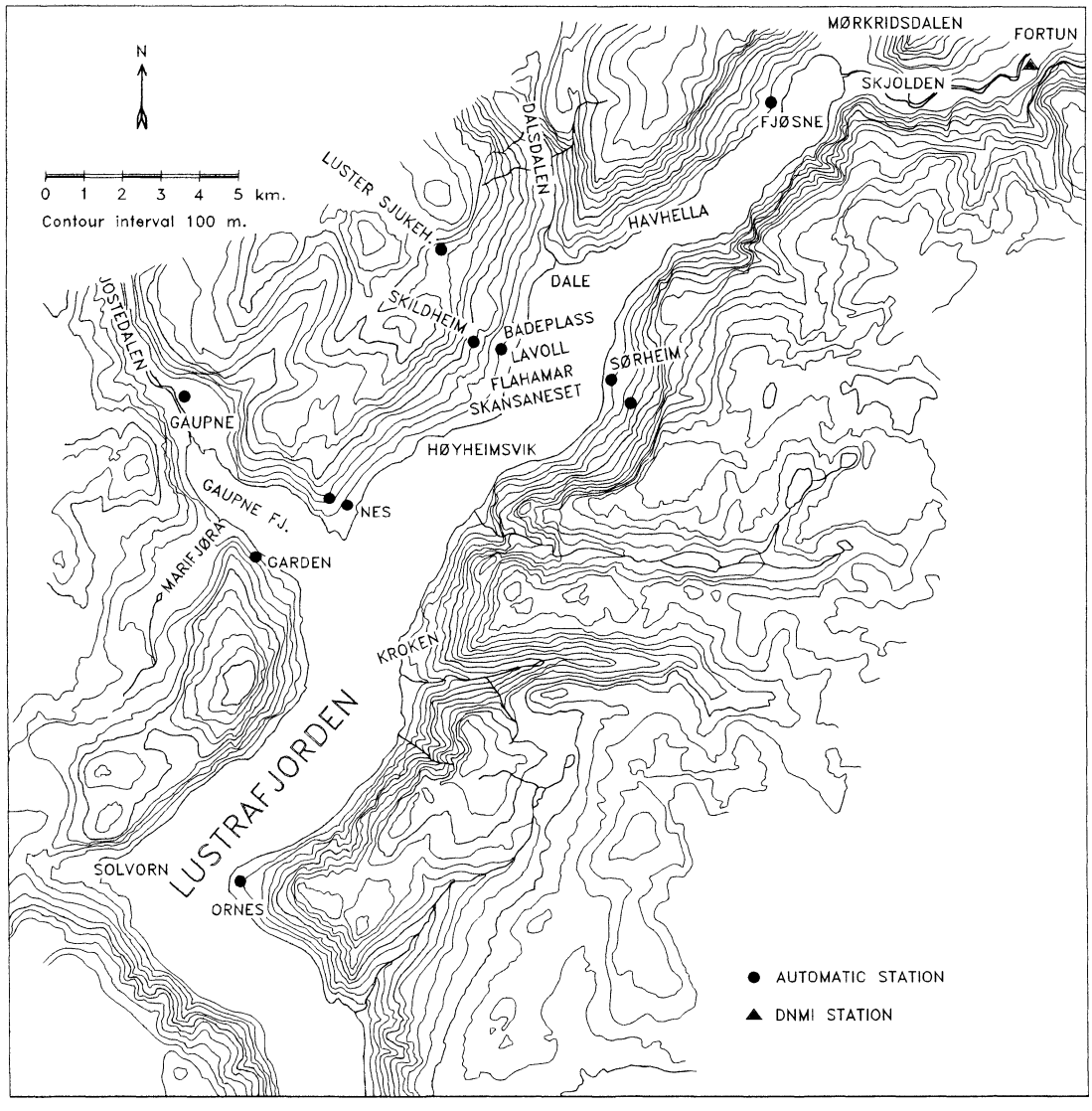


*Figure 3.31. Map of the Lustrafjord with climatic stations. At Fjøsne, Lavoll, Sørheim, Nes and Ornes air temperature were recorded at several altitudes above the fjord, 1981-85 (Hammer, 1986). The automatic stations (DNMI) Gaupne and Garden (established 1980), and Ornes, Skildheim, and Badeplass at Lavoll (established 1987) are still in operation (after Gjessing and Nordli, 1991).*

Freezing occurred nearly every winter, and practically simultaneously at the mouths of the rivers in Skjolden, and across the fjord from Dale to Sørheim (Figure 3.31). From Skjolden the sheet of ice expanded outwards the fjord, whereas that at Dale-Sørheim expanded both inwards and outwards. The thickness growth of the ice during the winter was, under otherwise similar circumstances, most pronounced when the ice had no snow cover. In general a fluctuation between fairly cold spells and mild, rainy spells that increased the fresh water flow to the fjord evidently provided the best conditions for the expansion of the ice sheet both in area and in thickness. Severe cold alone would rarely expand the area of ice markedly.

Besides ships - some sailing in fixed routes - strong winds were efficient icebreakers. When the ice field had gained a certain thickness it would withstand fairly long spells of rather mild weather in mid-winter. In late winter solar radiation contributed effectively to the weakening of the ice.

Then as now the main factors that prevented freezing of the fjord when the air temperature dropped below 0°C were special wind conditions, lack of precipitation, low disharge from rivers and brooks, and at times sailing ships.

The Gaupnefjord was not as easily iced over as the inner part of the Lustrafjord. Here freezing started at the mouth of the Jostedøla River and the sheet of ice spread towards the main fjord. The topographic, channeling effect on the wind is stronger, and more frequently evidenced in the Gaupnefjord than in the main fjord. This applies to drainage winds down the Jostedal Valley as well as the gradient wind which is often westerly."The Gåpna Wind" is the local term for these winds which for one thing are said to prevent - or at least restrict - the freezing of the fjord all the way towards Kroken. It should, however, also be noted that fjord steamers frequently called at Marifjøra, and their breaking up of newly formed ice together with "cooperative" winds, most certainly helped keeping the Gaupnefjord more or less ice-free.

***Fortun Power Plant*** (discharge in winter: mean 14 $m^3$ $s^{-1}$, maximum 25.5 $m^3$ $s^{-1}$) was put into operation in 1959. Boe and Roen (1991) assessed subsequent changes in ice conditions of the Lustrafjord:

- The prerequisite of precipitation, or large flow of fresh water from the catchment area prior to freezing was no longer neccessary, although it still typified the most favourable structure of the surface layer for a more or less spontaneous freeze-up of a wide area. Under otherways similar circumstances freezing of the inner part of the Lustrafjord occurred more frequntly than before the regulation.

- The continuous discharge of fresh water to the fjord also favoured the expansion, both in thickness and area, of the spontaneusly formed sheet of ice, which accordingly became less liable to break-up during mild spells.

- The fact that the temperature of water discharged from the power station was above 0°C when it poured into the fjord, and possibly partly also mixing with saline fjord water, induced a lead from the mouth of the Fortun River towards Fjøsne. The area of the lead varied mainly with the air temperature.

- Measurements of currents and salinity in the Lustrafjord support allegations of more frequent freeze-up also further outwards the fjord after than before regulation. This especially applies to the area between Flahamar and Nes, and under favourable circumstances also further outwards.

- The Fortun regulation had apparently minor effects on the ice conditions of the Gaupnefjord. In rare cases it may have caused some icing at the outher part of the fjord.

A marked reduction of the frequency of fjord steamers calling in the fjord around the end of 1964 enhanced the incidence of freezing and the duration of an established ice cover. However, it is not possible to quantify that effect.

***Leirdøla Power Plant*** (discharge in winter: mean 14 $m^3$ $s^{-1}$, maximum 25 $m^3$ $s^{-1}$) was started running in the autumn of 1978. Boe and Roen (op. cit.) found the subsequent changes in ice conditions:

- More frequent freeze-up of the main fjord from the mouth of the Gaupnefjord towards Solvorn-Ornes. Measurements of currents, temperature, and salinity indicated that under specific wind conditions icing also occurred more frequently inwards the fjord from Nes.

- When Leirdøla Power Station was running, extensive ice in the Gaupnefjord formed only at air temperatures below -15°C in mid-winter, and, apparently more easily in late winter. By more moderate cold, some ice formed in the bay between the outlet of Jostedøla and Marifjøra only.

- When the power station was stopped during cold spells, most of the Gaupnefjord froze up nearly spontaneously, and the ice grew rapidly in thickness. When the station anew was put into operation, a lead from the mouth of the river formed rapidly. Depending on the thickness of the ice and air temperature, the lead was gradually enlarged, and except by severe cold, most of the Gaupnefjord was cleared of ice. In mild weather full capacity running of Leirdøla Power Plant could even cause melting of the ice some way out into the main fjord. However, boats sailing in the area Solvorn-Gaupne-Høyheimsvik - and even more so an icebreaker occasionally hired by the State Power Company (the owner of Leirdøla and Jostedøla Power Stations) - and wind more frequently than disharge from the power plant resulted in the breaking up of the ice in this area.

***Jostedøla Power Station*** was started on 27.11.89. Pooled with the running water from Leirdøla Power Station the discharge from this station is channelled to a deep of some 40 m in the Gaupnefjord. Thus, mixing of the fresh water with huge quantities of saline, fjord water effectively prevents the formation of a stable surface

layer of brackish water. This also implies that the effect of the Leirdøla regulation on the ice conditions was removed.

### *3.5.3. Effects of ice versus open water on ambient air temperatur*

Strandenes (1984) studied the climatic effects of ice-covered versus open water at Lake Mjøsa, and found that at air temperatures lower than -15°C the ice cover on the average depressed the temperature by 4.4°C at Kise (5 m above Lake Mjøsa), and 3.0°C at Staur (30 m above Lake Mjøsa). The effect gradually decreased by increasing temperature. At Sand (3 m above Lake Storsjøen, see Figure 3.28) Pleym (1980) partly found somewhat larger effects.

Nordli (1981) studied the effects of ice on Nordfjord and on the adjacent Lake Loenvatn on ambient air temperature. The average effects on a station situated 3 m above the fjord is depicted in Figure 3.33. He also found that the ice caused a corresponding or somewhat greater depression of the air temperature at two stations 3 and 10 m above Lake Loenvatn respectively, the effect being most pronounced at the lower station.

In the ***Study of climate and frost damage in Luster*** air temperatures were recorded in 5 vertical sections along the Lustrafjord during the periode 1982-86 (cf. Figure 3.31). These data and data from weather stations operated by DNMI, and ice observations were analysed by Hammer (1986). Figure 3.32 shows temperature differences (mean differences and their standard deviations in various temperature groups) between Høyheimsvik (20 m a.s.l., in 1966 5 m a.s.l.) and the reference station Fortun by ice-covered and ice-free fjord. The standard deviations by ice-covered fjord for the winters 1981/82-1984/85 indicate very great variations in the temperature differences, especially in the lower temperature groups. This is probably due to differences in wind conditions between Fortun and the fjord area. By low temperatures, calm or light winds prevail at Fortun. Simultaneously the fjord area may

experience moderate winds. The accumulation of stagnant cold air above the ice field in the fjord area is particularly strong by a weak gradient wind up the fjord. Under these circumstances the temperature difference between Høyheimsvik and Fortun is usually small, and occasionally negative. With increasing wind down the fjord and/or marked drainage winds down the sides of the fjord the positive difference increases.

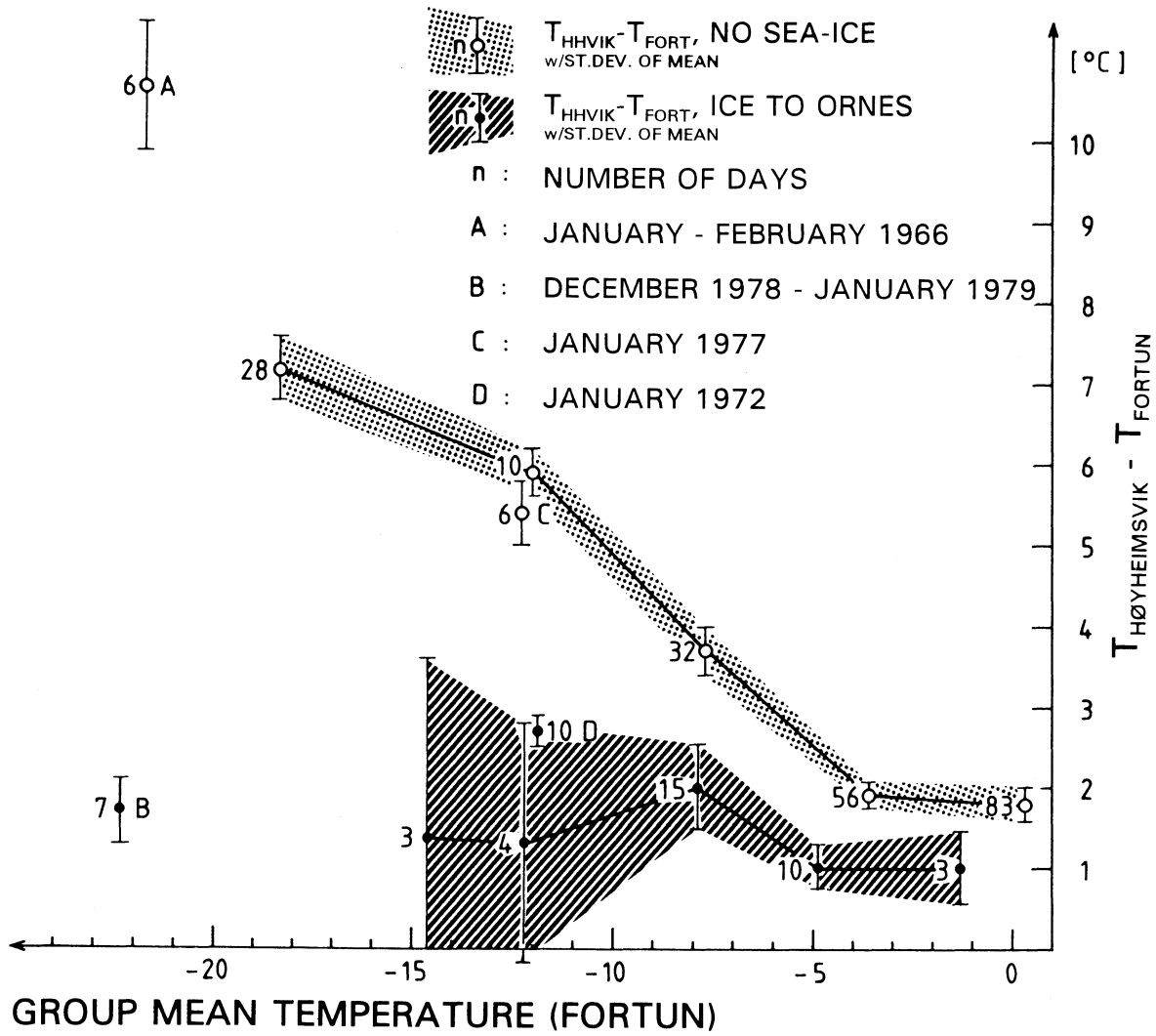


*Figure 3.32. Temperature differences between Høyheimsvik and Fortun - in various temperature groups referred to Fortun - at ice-covered and ice-free fjord. Group means and their standard deviations for the winters 1981/82-1984/85 are shaded (after Hammer, 1986).*

The extremely large differences by severe cold in 1966 (A in Figure 3.32) are partly due to the fact that the site of the station Høyheimsvik (5 m a.s.l.) then was more strongly influenced by airflows from the ice-free fjord than the subsequent site (20 m a.s.l.).

By means of the temperature difference method (Appendix 6.1) - with Fortun as reference station - Hammer calculated average effects of an ice cover (3 groups: The inner part of the fjord frozen over: I. to Ornes, II. at least to Nes, and III. at least to Dale-Sørheim) versus open water on the air temperature at all sites of observation. In Figure 3.33 is given a smoothed curve depicting the effect in the shore zone in the central parts of the fjord when the ice field reaches beyond Nes. Since there were few cases with temperatures below -20°C, the results at very low temperatures are uncertain. However, the tendency to increasing effect with decreasing temperature is well documented. It is also noted that the effects of an ice cover on air temperature in the shore zone in Luster and in the inner part of Nordfjord nearly coincide.

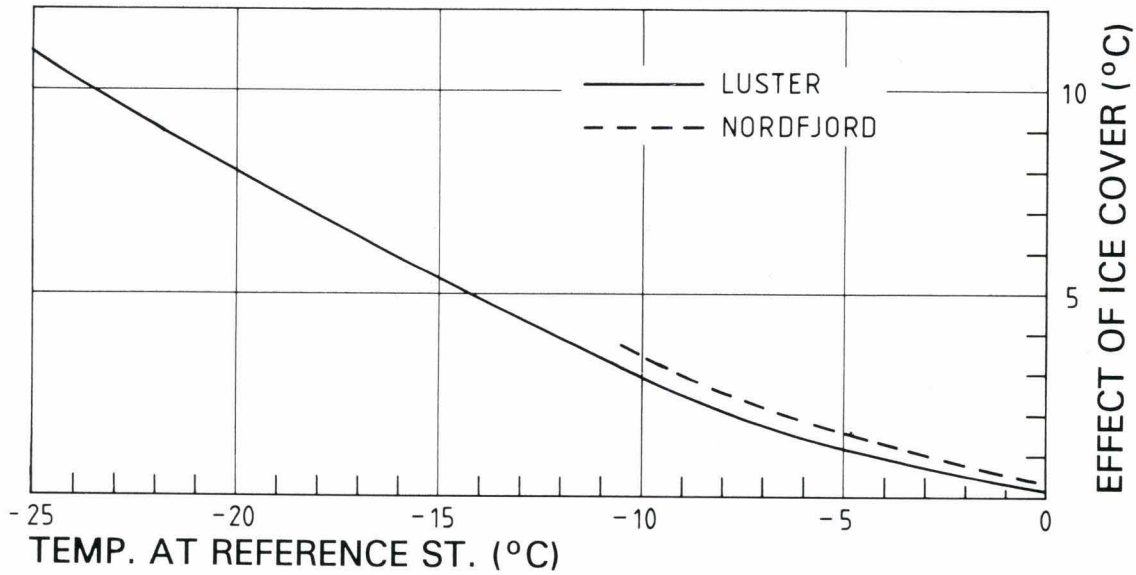


*Figure 3.33. Smoothened effect of an ice cover on air temperature in the shore zone of Lustrafjord and at the inner part of Nordfjord (after Hammer, 1986 and Nordli, 1981).*

Hammer (op. cit.) also summarized the effect of the ice-covered Lustrafjord on the air temperature at different altitudes on the sides of the fjord in a table which is not reproduced here.

Based on the results presented by Hammer (op. cit.) and extensive new recordings (cf. Figure 3.31) Gjessing and Nordli (1991) developed a method for calculating the effect of the regulations - i.e. of the induced changes in ice conditions - on air temperatures along the sides of the Lustrafjord.

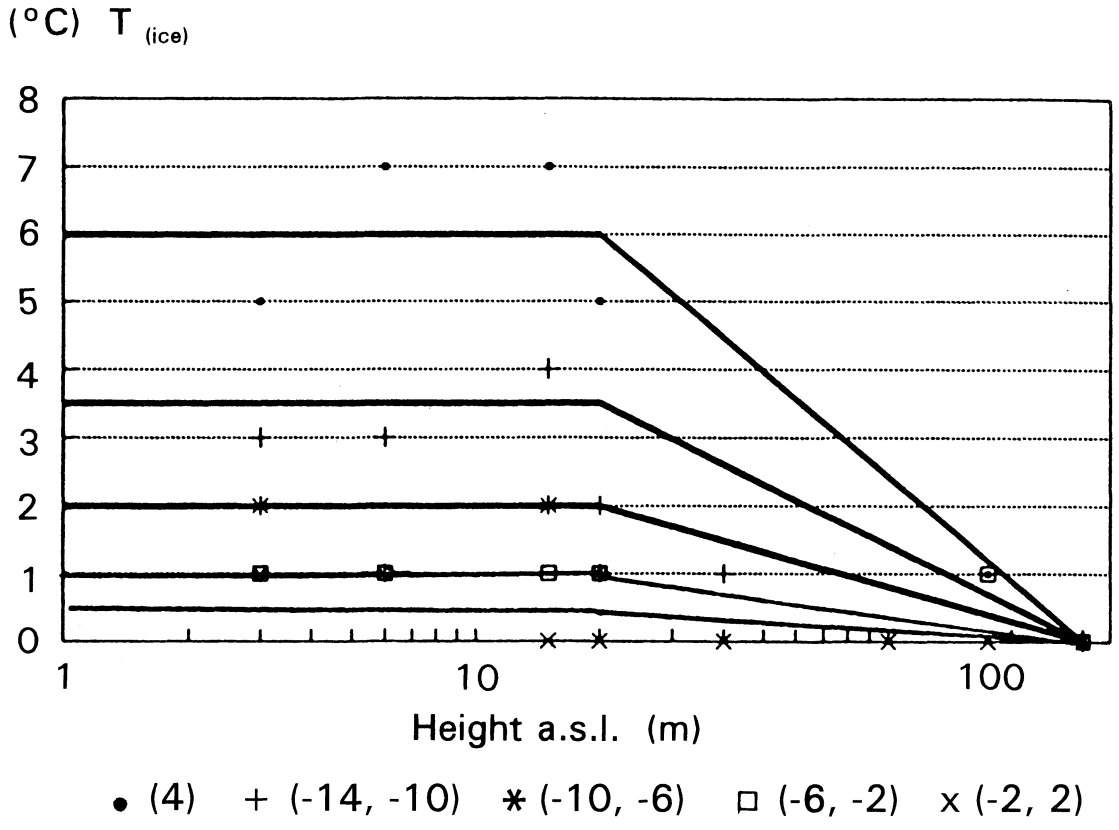


*Figure 3.34. Smoothened effect of an ice cover on the air temperature in a vertical section upslope the sides of the Lustrafjord based on recordings from Lavoll, Høyheimsvik, and Skildheim. The data was grouped in 5 temperature classes based on recordings from the reference station Fortun (after Gjessing and Nordli, 1991).*

Figure 3.34 is based on recordings when the fjord was frozen up down to, or beyond Nes. At the lowest 20 m of the sides of the fjord there was evidently no marked diminishing effect of the ice with increasing altitude. Further upslope the effect obviously decreased - first rather rapidly, and, then more slowly - up to about 300 m a.s.l. Gjessing and Nordli found a representative model by assuming a constant effect in the lower layer, and a logarithmically decreasing effect with altitude in the upper layer, as shown in Figure 3.34 for 5 selected temperature classes.

The graphic representation in Figure 3.34 is expressed by the equations:

$$\bar{E}_{S,ice} = K \qquad for\ h < h_1 \qquad (3.2)$$

*and*

$$\bar{E}_{S,ice} = \frac{\log h_2 - \log h}{\log h_2 - \log h_1}\ (°C) \qquad for\ h_1 \leq h \leq h_2 \qquad (3.3)$$

where $h$ is altitude (m a.s.l.), $K$ (°C) is the effect of the ice in the lower layer, and $h_1$, $h_2$ are 20 and 300 m respectively. $K$ is a function of local terrain forms (straight shoreline or headland) and air temperature at the reference station Fortun. Values of $K$ for fjord sides with straight shorelines can be read from Figure 3.34, or from Table 3.8. Evaluation of data from Nes and Ornes when the fjord was frozen up beyond those headlands, yielded similar results, but the $K$-values were somewhat greater (Table 3.8).

By ***ice-free fjord*** Gjessing and Nordli found the average differences in air temperature between Fortun and localities in the shore zone of the Lustrafjord presented in Table 3.9, and in addition a mean vertical temperature gradient of - 0.5°C per 100 m. From recorded temperatures at Fortun they could now calculate average temperatures at any locality at straight fjord sides and headlands both by ice-covered and ice-free fjord.

*Table 3.8. Values of K (°C) in equation (3.2) as a function of air temperature at Fortun, T , and local form of the shoreline (after Gjessing and Nordli, 1991).*

| Temperature interval | Straight shoreline | Headland |
|---|---|---|
| -2 < T ≤ 2 | 0.5 | 1.0 |
| -6 < T ≤ -2 | 1.0 | 2.0 |
| -10 < T ≤ -6 | 2.0 | 3.5 |
| -14 < T ≤ -10 | 3.5 | 6.0 |
| T ≤ -14 | 6.0 | 8.0 |

*Table 3.9. Mean differences in air temperature, $A_0$ (°C), between Fortun and the shore zone of the Lustrafjord by ice-free fjord, for localities on straight shorelines and headlands respectively* (after Gjessing and Nordli, 1991).

| Temperature interval | Straight shoreline | Headland |
|---|---|---|
| -2 < T ≤ 2 | -1.2 | -1.8 |
| -6 < T ≤ -2 | -2.1 | -3.1 |
| -10 < T ≤ -6 | -3.9 | -4.5 |
| -14 < T ≤ -10 | -5.0 | -6.5 |
| T ≤ -14 | -6.5 | -8.0 |

By ***ice-covered fjord*** the following equation is valid:

$$T_{S,h} = T_F - A_0 - 0{,}005°C \cdot h - \bar{E}_{S,ice} \qquad (3.4)$$

where $T_{S,h}$ (°C) is the temperature at site S at altitude h (m), $T_F$ is the temperature at Fortun, $A_0$ is the appropriate temperature difference given in Table 3.9, and $\bar{E}_{S,ice}$ is the effect of the ice cover calculated by means of eqn. (3.3). The corresponding temperature by ***ice-free fjord*** is obtained by putting $\bar{E}_{S,ice} = 0$.

Ice conditions were, as referred above, studied by Boe and Roen (1991). They also computerized the data. For various parts of the fjord and for a series of winters the ice cover was classified on a scale from 0 to 4 (0: ice-free, and 1: 25%, 2: 50%, 3: 75%, and 4: 100% ice cover respectively), and the effect of the regulations on ice conditions was estimated. Gjessing and Nordli (op. cit.) applied these data in their model, and calculated various temperature parametres - among others minimum temperatures, and degree days in early spring.

*Table 3.10. Examples from some winters showing the effect of the regulation on the lowest minimum temperature, $T_{N,a}$, and the number of days with minimum temperature below certain thresholds (°C) 10 m a.s.l. at exposed terrain (small headlands) in the area Dale - Nes (after Gjessing and Nordli, 1991).*

| Winter | State | $T_{N,a}$ | Number of days with $T_N$ equal to or below | | | | |
|---|---|---|---|---|---|---|---|
| | | | -25 | -20 | -15 | -10 | -5 |
| 71/72 | Reg. | -15.3 | | | 1 | 22 | 59 |
| " | Nonreg. | -15.3 | | | 1 | 22 | 59 |
| 72/73 | Reg. | -13.6 | | | | 5 | 15 |
| " | Nonreg. | -7.6 | | | | | 9 |
| 76/77 | Reg. | -14.4 | | | | 12 | 54 |
| " | Nonreg. | -13.5 | | | | 7 | 51 |
| 78/79 | Reg. | -26.4 | 2 | 9 | 34 | 67 | 86 |
| " | Nonreg. | -22.5 | | 5 | 28 | 54 | 81 |
| 80/81 | Reg. | -22.5 | | 2 | 21 | 42 | 69 |
| " | Nonreg. | -22.5 | | 2 | 21 | 42 | 69 |
| 82/83 | Reg. | -14.5 | | | | 5 | 22 |
| " | Nonreg. | -6.4 | | | | | 9 |

The calculations revealed no significant effects of the regulations on the start of the growing season or on accumulated degree days in spring.

Table 3.10 presents some simplified, selected examples of results arrived at for the minimum temperature. It is evident that the largest changes - in the lowest winter temperature, $T_{N,a}$, as well as in the number of days with low temperatures - due to regulation ocurred in moderately mild winters as e.g. in 72/73 (only the Fortun/Granfasta regulation) and in 82/83 when also Leirdøla Power Station was in operation. However, also in the exceptionally severe winter of 78/79 the assessed changes due to the regulations were large, whereas no changes caused by the regulations were evident in the severe winter of 80/81. The latter result indicates that the old local saying: "If the fjord is not already partly or entirely frozen up, it will never freeze during severe cold", is valid also subsequent to regulation.

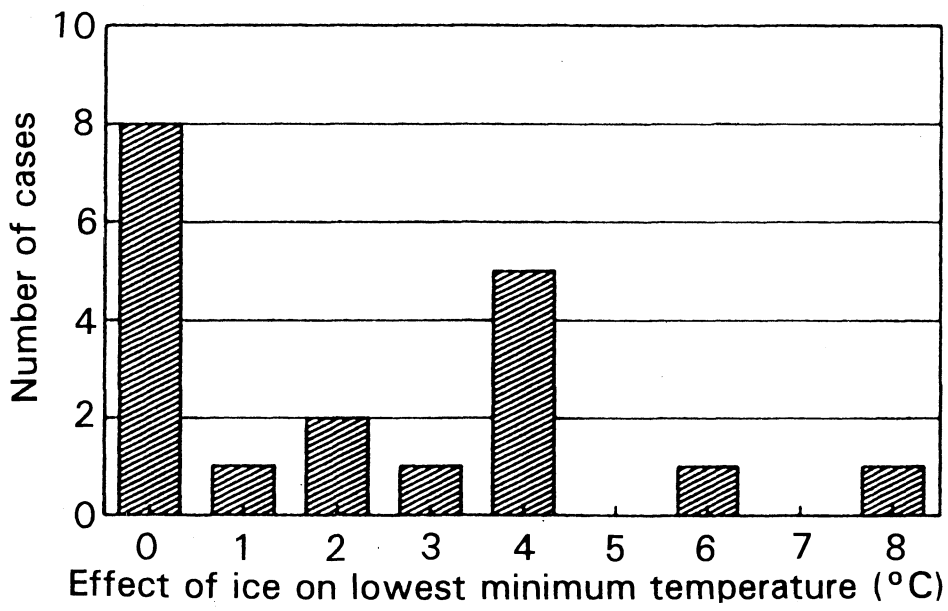


*Figure 3.35. The effect of the regulations on the lowest minimum temperature at well exposed sites near the fjord in the area Dale-Nes during the winters 1970/71-1988/89 (after Gjessing and Nordli, 1991).*

The results for the lowest minimum temperature during the winter for the entire periode 1970-1989 are depicted in Figure 3.35. It is seen that in 11 out of 19 winters the regulations have caused lower minimum temperature in this area.

The studies in the Lustrafjord clearly evidenced that the changes in ambient air temperature due to the regulations are caused by the induced changes in ice conditions of the fjord. The effect of the ice cover is most pronounced by low temperatures and calm or light winds, and it decreases with altitude above, and distance from the fjord. Under these circumstances - particularly by ice-free fjord - cold air drains down the sides of the fjord and down valleys leading to the fjord. Accordingly, the heat flow from open water towards those areas is more or less impeded (cf. Figure 2.3). For example, in Gaupne at the mouth of the Jostedal Valley continuous recordings since 1980 have revealed no effects of ice-covered versus open water on the air temperature.

On steep sides along the fjord, and especially at headlands and on small hills, the air flow from higher levels is usually greatly reduced, and therefore an onshore airflow from open water may more easily invade the lower levels of such areas. For the lowest temperatures Gjessing and Nordli showed that the average effect of ice versus open water on the air temperature in the shore zone was some 6°C on steep slopes, and some 8°C at headlands (Table 3.8).

Briefly summarized, Gjessing and Nordli (op. cit.) found that the regulations have had little or no effect on ambient air temperature in the area from Skjolden to Dale-Sørheim, where changes in ice conditions due to the regulations are small. For the area Dale-Sørheim to Nes - Kroken they showed that in 11 out of 19 winters the regulations had induced lower minimum temperatures in the shore zone (Figure 3.35). The temperature depression was mostly 1-4°C, the largest drop at well exposed sites being some 8°C. The area further down the fjord to Ornes is more rarely affected by the regulations - especially subsequent to the construction of the deep-water outlet from Leirdøla and Jostedøla Power Stations.

However, if the fjord freezes up here, the minimum temperature may drop some 4-8°C.

A possible lag in the break-up of the ice in late winter or early spring due to the regulations, has obviously no significant effect on calculated temperature-growth indices (degree-days). This indicates that the regulations have not affected the start of the growing season at any localities along the fjord.

### *3.5.4. Effects on the humidity of the air*

Changes in ice conditions in a fjord due to regulation may also affect the humidity of ambient air. When the fjord is frozen up, the lead at the mouth of a river pouring discharged water from a power station into the fjord is much wider than in the natural state. During cold spells this augments the transfer of water vapour to the air, and by calm or light winds and severe cold, frostsmoke may develop, and heavy hoarfrost deposition may occur on surfaces in the vicinity. By sudden clearing of the sky and calm winds a rapid drop in the air temperature frequently produces radiation fog above the ice cover of a fjord. Slowly advected moisture from a wide lead at the mouth of a river will enhance the development of such fog. Radiative loss from the top of the fog layer cools the entire layer making the fog more persistent. However, since the saturation vapour pressure at an ice surface is lower than at a water surface of the same temperature (Table A1, Appendix), continuous deposition of hoarfrost will take place from fog consisting of water droplets, i.e from fog at a temperature of some -20°C or higher. Therefore, such fog dissipates near the surface, but a low stratus may persist. Hammer (1986) observed rather heavy hoarfrost deposit on vegetation and other obstacles which had been enveloped in such fog. Except in such situations he mostly found only small deposits of hoarfrost along the Lustrafjord. It should be noted that also when a fjord is frozen up, evaporation from leads at the mouth of regulated rivers supplies only part of the moisture in ambient air (cf. Chapter 3.4.5).

Hammer (op. cit.) found that by ice-free fjord and severe cold, frostsmoke frequently formed at the open Fortundal River (downstream of Fortun Power Station), but rarely at Jostedøla River (downstream of Leirdøla Power Station). The main reason for this difference was a weaker mountain wind in the Fortun Valley than in the Jostedal Valley. On rare occasions ice covering the Gaupnefjord and the adjacent part of the Lustrafjord would favour local accumulation of cold air, which inhibited the drainage wind at the mouth of the Jostedal Valley. Then extensive frost smoke could develop downstream of Leirdøla Power Station.

# 4. POSSIBLE DETRIMENTAL EFFECTS OF INCREASED AIR HUMIDITY, FROSTSMOKE, AND HOARFROST

## 4.1. Introduction

Claims have been made that frostsmoke and heavy hoarfrost deposits in meadows and pastures may induce compact icing, which may cause "iceburn", i.e. in spring patches or wide areas appear as scorched by the sun. It is, however, difficult to find a plausible, physical explantion for such an icing process. Utaaker and Jackhelln (1965) showed that changing weather (mild weather with rain, or sleet on frozen ground, followed by severe cold) is the likely cause of icing in meadows.

It has also been claimed that during cold spells increased humidity of the air may affect harmfully the moisture conditions in houses for domestic animals. Moreover, it has been suggested that humid air and frost smoke can invade storehouses and cause heavy deposit of hoarfrost e.g. on cured meat or on hay, which incidentally may go bad. The problems concerning conditions in cowhouses and haybarns are discussed below (Chapters 4.2 and 4.3).

To people who live, work, or travel in the area an increase in humidity, frostsmoke and hoarfrost may be disagreeable. In some affected areas local fog may slow down road traffic at times, but it is rarly hazardous. Otherwise, it is rather difficult to assess the consequenses for the well-being of people and animals. Apparently, direct harmfull effects on human or animal health have not been documented in scientific literature.

Neither has it been possible to find evidence of injuries to plants (trees and scrubs) due to heavy deposit of hoarfrost. Heavy accretion of rime onto trees, power lines etc. may cause mechanical damage (breakage), mainly when super-cooled clouds envelop and move across exposed sites. Rime is rarely deposited from frostsmoke.

Falling snow will hardly cling more onto branches covered by hoarfrost than to trees with no frost, whereas branches covered by rime may accumulate a more heavy load of snow.

Condensation of water vapour in fuel tanks etc. occurs mainly by rapid rise in temperature and humidity of the air, i.e. by sudden changes from cold to mild, humid weather. Under these circumstances locally added water vapour from an open river does not give a measurable contribution to the moisture content of the air.

Assessment of effects on materials and paint due to increased deposit of hoarfrost on buildings is also missing in scientific literature. This also applies to problems concerning increased risk of rusting and corrosion of metals.

## *4.2. Moisture in houses for domestic animals*

The optimal indoor climate differs for various species of domestic animals. Lilleng and Merok (1984) conclude that *air temperature* is the main climatic element, and that *humidity of the air* seems to be of less importance. However, most animals reacts against extreme conditions, as e.g. coinciding high temperature and high relative humidity. High humidities giving rise to heavy condensation may also deteriorate hygienic conditions and cause damage to building constructions.

For example, optimal air temperature for dairy cattle is 10-12°C, and for sows with piglets 15-20°C, whereas relative humidity should be between 70 and 80%. In a climatically conditioned cowhouse total exchange of air twice per hour is a rule of thumb. By outdoor, low freezing temperatures ventilation must often be somewhat reduced, since heat originating from the animals (the metabolism) on the average is fairly constant.

During calm, cold weather in winter regulation induces an increase in absolute humidity of the air along parts of the river where there has been a marked increase in the area of open water (cf. Chapter 3.4.3). However, in most cases a simultaneous rise in air temperature increases the "drying power" (the water vapour defisit) of air that invades buildings (Utaaker and Skaar, 1986, Table 3.9, p. 57).

Extreme effects of increased moisture in outdoor air on the humidity of the air in a cowhouse is illustrated here by an example where the rise in air temperature is not taken into account:

Prior to regulation outdoor air temperature, $T_a$ = -20°C, and relative humidity, RH = 44%, i.e. the absolute humidity, $\rho$ = 0.47 g $m^{-3}$. In a similar situation after regulation $T_a$ = -20°C, whereas RH is increased to 100%, i.e $\rho$ = 1.07 g $m^{-3}$. In the cowhouse $T_a$ = 10°C and RH = 75%, i.e, $\rho$ = 7.05 g $m^{-3}$. By total ventilation twice per hour 1 $m^3$ outdoor air is on the average mixed with 29 $m^3$ indoor air per minute. An increase in $\rho$ of 0.6 g $m^{-3}$ in the ventilative air will increase the RH in the cowhouse by some 0.17%. In 30 minutes equilibrium is reached, and the RH is increased by some 6%. Considering the possibility that in an extreme frostsmoke situation droplets amounting to 0.2 g $m^{-3}$ are added to the fresh air. Then the total increase in moisture content is 0.8 g $m^{-3}$, and and the corresponding equilibrium RH is some 8.5% percent above that prior to regulation.

In general, frostsmoke dissipates before the air enters the fresh air intake, and RH of this air is close to equilibrium saturation above ice (cf. Table A1, Appendix). The recordings in the Orkdal Valley (Utaaker og Skaar, 1986) also showed that e.g. at $T_a$ = -20°C RH prior to regulation was rarely below 60% at any of the valley stations. The actual increase in RH in a cowhouse due to regulation is therefore always less than in the example above.

The mode of transfer of water vapour and sensible heat in the atmosphere is approximately similar. If an open water surface has caused an icrease in the absolute humidity of air entering a

cowhouse, the air is also heated (cf. Appendix 6.2.1, and Chapter 3.4.3). By low freezing temperatures part of the added water vapour is deposited as hoarfrost onto the ground and onto obstacles exposed to the air flow. The release of latent heat of sublimation slightly heats the air. Therefore, normally the rise in air temperature is measurable at a greater distance from open water than the simultaneous rise in moisture content due to regulation.

The temperature rise of the fresh air as well as its increased moisture content may affect the relative humidity in a ventilated cowhouse. An example from a somwhat more real situation than that discussed above, shows in detail how the effect can be calculated:

***Before regulation***: $T_a$ = -15°C and RH = 62.5%, i.e. $\rho$ = 100 g $m^{-3}$ in the outdoor air ($A_0$). It is further assumed that at a rate of total air exchange twice per hour the air conditioner kept $T_a$ = 10°C and RH = 75%, i.e. $\rho$ = 7.05 g $m^{-3}$ in the indoor air ($A_i$). Then the budget for absolute humidity of the air mixture in the cowhouse can be written :

$$\frac{\rho \text{ in } 1\ m^3 A_o + \rho \text{ in } 29\ m^3 A_i + \text{vapour from animals}}{30\ m^3} = \rho \text{ in } A_i \qquad (3.5)$$

i.e.

$$\frac{1.00\ g \times 1 + 7.05\ g \times 29 + 6.05\ g}{30\ m^3} = 7.05\ g\ m^{-3} \qquad (3.6)$$

The budget for air temperature is:

$$\frac{T_a \text{ in } 1\ m^3 A_o + T_a \text{ in } 29\ m^3 A_i}{30\ m^3} + \begin{matrix}\text{Temp.rise} \\ \text{(heat from anm.} \\ \text{+ heating)}\end{matrix} = T_a \text{ in } A_i \qquad (3.7)$$

i.e.

$$\frac{-15°C \times 1 + 10°C \times 29}{30} + 0.83°C = 10°C \qquad (3.8)$$

***After regulation:*** In an analogous situation, the properties of the outdoor air is modified by the open river to $T_a$ = - 13.5°C and RH = 87%, i.e. $\rho$ = 1.50 g $m^{-3}$. (Note that the rise in temperature corresponds to that found by Utaaker and Skaar, 1986, whereas the increase in absolute humidity is about twice the value they found). If the air conditioner is run like prior to regulation, and the contribution of water vapour and heat from the animals is unchanged, the following budgets apply:

For absolute humidity:

$$\frac{1.50\ g \times 1 + 7.05\ g \times 29 + 6.05\ g}{30\ m^3} = 7.067\ g\ m^{-3} \qquad (3.9)$$

For temperature:

$$\frac{-13.5°C \times 1 + 10°C \times 29}{30} + 0.83°C = 10.05°C \qquad (3.10)$$

In saturated air at 10.05°C the absolute humidity is 9.43 g $m^{-3}$, and the relative humidity in the cowhouse is:

$$RH = \frac{7.067}{9.43} \times 100\% = 74.94\% \qquad (3.11)$$

This represents conditions after 1 minute of ventilation. When equilibrium is reached after 30 minutes, the absolute humidity of the indoor air is increased to 7.55 g $m^{-3}$ and the temperature is raised to 11.5°C.

In saturated air at 11.5°C the absolute humidity is 10.34 g $m^{-3}$, and the new, stable relative humidity is:

$$RH = \frac{7.55}{10.34} \times 100\% = 73\% \qquad (3.12)$$

In this example the relative humidity of the air in the cowhouse has been lowered from 75% to 73% due to regulation. If $A_0$ after regulation had RH = 95%, corresponding equilibrium RH would be 74.9%, and with RH = 100% in $A_0$, it would be 76%.

If conditions in the cowhouse prior to regulation were e.g. $T_a$ = 15°C, and RH = 75%, corresponding calculations yield equilibrium RH equal to 72, 73.3, and 74% respectively, and $T_a$ = 16.5°C after regulation.

These calculations indicate that in a cowhouse, which after regulation, in winter is subjected to an air flow from open water, indoor climate is not harmfully affected by the changed environment. Provided otherwise unchanged conditions, indoor relative humidity has as a rule dropped somewhat due to regulation, whereas indoor air temperature has risen.

Lilleng and Merok (1984, p. 5) write: " There is a number of reasons for regular air exchange in animal houses. In winter the exchange must be vigorous in order to get rid of the water vapour produced by the animals, and to bring the content of repugnant gasses and pollutants down to acceptable concentrations. As a rule of thumb the minimum exchange should keep RH below 80%." Further they write, p. 9: " Normally animal houses are dehumidified by ventilation. Frequently, however, the flow of sensible heat from the animals is too small to allow a ventilation that brings RH down to an acceptable level, i.e. some 80%. Better insulation of the rooms and/or artificial heating are possible solutions of this problem." This implies that the rise in the temperature of ventilated animal houses, brought about by regulations, has hardly any detrimental effects on the indoor environment on cold winter days.

Occasionally "fog" is observed in fresh air entering an animal house. This is not frostsmoke from outside air, but typical interior ***mixing fog*** (cf. Figure 2.5). For example, if 1 $m^3$ fresh air ($T_a$ = -20°C and RH = 85%, i.e. $\rho$ = 0.91 g $m^{-3}$) is mixed with 1 $m^3$ indoor air ($T_a$ = 10°C and RH = 75%, i.e. $\rho$ = 7.05 g $m^{-3}$) the mixture ($T_a$ = -5°C and $\rho$ = 3.955 g $m^{-3}$) is supersaturated (cf. Appendix, Table A1), and fog droplets form instantaneously. When this "parcel" of air is further mixed with indoor air the fog will dissipate rapidly. In the present example a mixing ratio (fresh air)÷(indoor air) of 1÷2.3 would give ***just no fog***, and with 0.5 g $m^{-1}$ less moisture, i.e. RH = 38%, in the fresh air this ratio would be 1÷1.9. This indicates that even when the outdoor air is rather dry, "fog" may form at the fresh air intake in an animal house. However, since an increased area of ice-free water causes a rise both in temperature and absolute humidity of ambient air, fog formation at the fresh air intake of animal houses in such localities is hardly influenced by the regulation.

Neither can the small increase in absolute humidity of fresh air contribute measurably to possible condensation onto indoor walls and ceilings. Compared to the water vapour released from the animals it is insignificant. Condensation, at low outdoor temperatures, in an animal house with appropriately dimensioned and correctly run ventilation, is mainly due to insufficient insulation of walls and/or ceiling. If the temperature of a surface drops below the dewpoint temperature, $T_D$ , of the air, water vapour "proceeds" towards "the cold wall" and condenses onto it as dew. With $T_a$ = 10°C and RH = 75%, $T_D$ = 5.5°C, i.e. dew will form on "the cold wall" when its temperature drops slightly below 5.5°C. With $T_a$ = 10°C and RH =80%, $T_D$ = 6.5°C, and with RH = 90%, $T_D$ = 8.4°C. Corresponding values for $T_a$ = 15°C, and RH equal to 75% , 80%, and 90% are $T_D$ equal to 9.4°C, 10.4 °C, and 13.3 °C respectively.

## *4.3. Moisture in a hay barn*

*Figure 4.1. The barn on the small island in Orkla where the experiment was conducted (photo E. Skaar).*

To study possible effects of changes in outdoor climate on the moisture content of hay stored in barn, measurements were conducted through the winters of 1980/81-1984/85. In a small barn close to the River Orkla (Figure 4.1) weight changes of a basket filled with some 160 kg dry hay (Figure 4.2), and a number of weather elements inside and outside the barn were recorded continuously (Utaaker and Skaar, 1987a,b).

Two examples of results of these recordings are given in Figure 4.3. - one from the period ***before*** and one for that ***after*** the regulations were implemented. The main results of the study are summarized in the following:

1. When the temperature of the hay was above 0°C, the hygroscopic properties of the hay played a major role in the moisture exchange with ambient air. When the hay was frozen, i.e. the surface temperature was below 0°C, hoarfrost deposition onto the hay or sublimation from the hay were the main processes.

2. During cold, calm spells there were no or only minute changes in the weight (the total moisture content) of the hay. The diurnal increase in weight was rarely more than 50-60 g,

*Figure 4.2. Part of the weighing equipment with the basket of hay inside the barn (photo E. Skaar).*

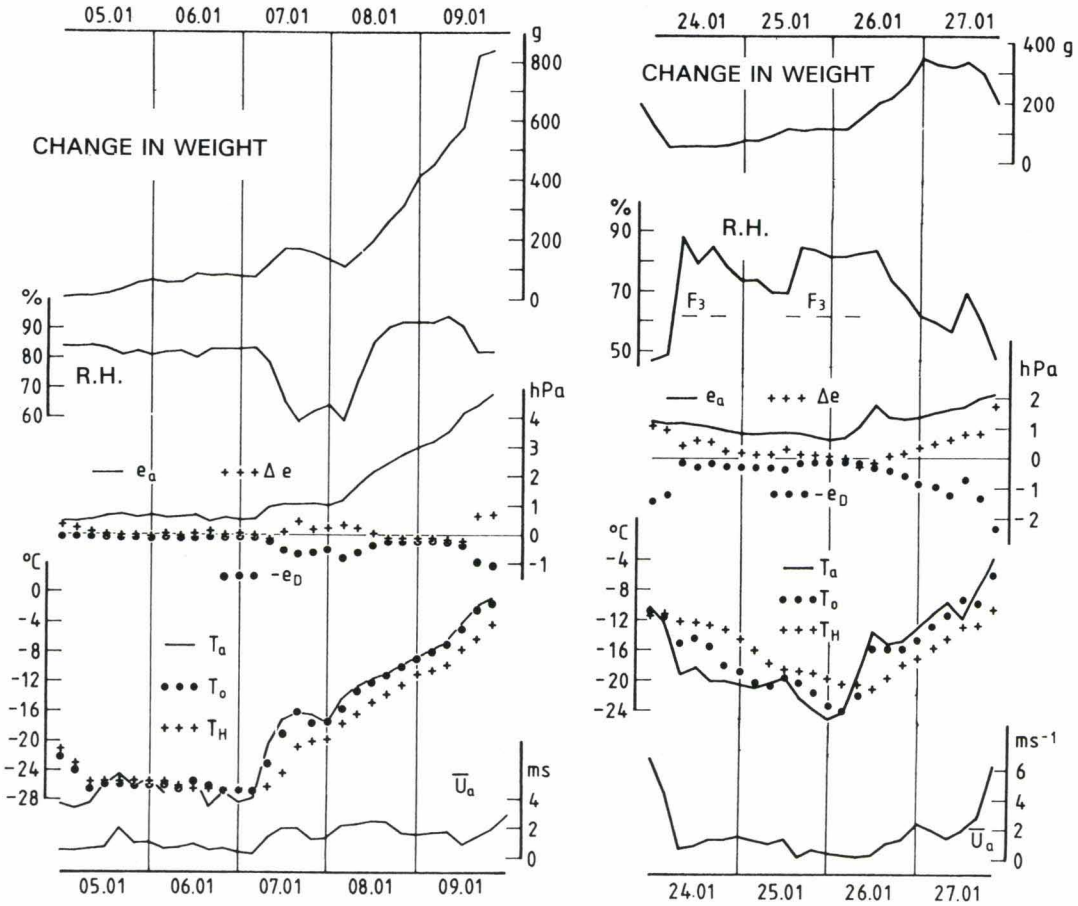


*Figure 4.3. Four-hourly averages of changes in the weight of the hay. RH = relative humidity, $e_a$ = water vapour pressure, $e_D$ = vapour pressure deficit, $\Delta e = e^*_0 - e_a$ (where $e^*_0$ is saturation vapour pressure over ice at temperature $T_0$), $T_a$ = temperature, and $U_a$ = wind speed of outdoor air. $T_0$ = surface temperature of the hay, $T_H$ = temperature within the hay.* ***To the left*** *for 5-9 January 1982, before Grana Power Station was put in operation.* ***To the right*** *for 24-27 January 1984, with the station in full operation. $F_3$ under the RH-curve on 24 and 25-26 January 1984 indicates dense frostsmoke around the barn (after Utaaker and Skaar, 1987a).*

corresponding to a frost deposit eqvuivalent of some 0.005 mm water onto the exposed surface of the hay. Both ***before*** and ***after*** regulation such increases always occured in connection with a small rise in the air temperature while the relative humidity of the air remained high. No significant changes caused by regulation appeared.

3. Even dense, extensive frostsmoke ***after*** regulation had no influence on the weight of the hay (cf. Figure 4.3).

4. ***Before*** as well as ***after*** the implementation of the regulations the largest increases in the weight of the hay took place by sudden changes from cold, calm weather to very mild weather with moderate or strong up-valley winds, i.e. when mild, humid, maritime air pushed away the cold air over the region. The largest diurnal increase, some 1.9 kg, was recorded in January 1981, i.e. ***before*** regulation. Then sublimation and condensation onto the hay corresponded to some 0.2 mm of water spread over the exposed surfaces of the basket. However, this was sufficient to make the surface layer of the hay rather wet. By mild, humid weather in winter evaporation from the open river is so small that the local contribution to the moisture of the air is negligible.

5. The largest reductions in the weight of the hay occured when moderate or strong down-valley winds caused a marked rise in temperature and a notable fall in relative humidity of the air. During spells of persistent low relative humidity and fresh winds, the rate of weight loss usually diminished gradually. The regulation caused no evident changes in these conditions.

This study evidences that the exchange of moisture between invading outdoor air and the dry hay in the barn was governed by the weather. Local frostsmoke and increased absolute humidity of the air caused by regulation had no measurable effects on the moisture content of the hay.

In a comprehensive study around the outlet of Lake Vågåvatn Nordli (1986) obtained similar results. He found no significant effects of frostsmoke on the weight of hay in "net bags" hung up in exposed barns in the area.

The conditions of hay stored in a barn, are never quite like these of the experiments referred to above. In general, dry, fresh hay undergoes a weight (water) loss during storage - partly due to microbiological activity. Especially during cold spells water vapour is transferred towards the cold surface. Under similar circumstances considerably more moisture would appear as hoarfrost on the surface of a several-metre-thick layer of hay than on the surface of the hay in the basket or the "bags" of the referred experiments. When the surface temperature rises above 0°C, and the frost thaws, the surface layer of the ordinarily stored hay will be much wetter than those of the the basket and the "bags". However, the physical laws governing the exchange of water vapour between hay and ambient air are everywhere the same. Hence, probably, the conclusions listed above are also applicable to ordinarily stored hay in barns close to rivers downstream of power stations. This also applies to other types of dry forage and food stored in fairly well ventilated houses in such areas.

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# 6. APPENDIKS

## 6.1. The difference method

The difference method (Lowry, 1976) may be applied e.g. to study effects of artificial or natural changes in the landscape on local air temperature:

The temperature, $T$ , at a site, $S$ , at the time, $t$ , can be expressed by:

$$T_{S,t} = C_{S,t} + L_{S,t} + E_{S,t} \qquad (1A)$$

where $C$ is the effect of macro-scale weather on the air temperature, $L$ are local effects of the of the landscape, and $E$ are effects due to changes in $L$.

The method requires a reference station, $R$ , that is not affected by changes in $L$ , but is situated so close to $S$ that macro-scale weather is nearly the same at the two sites.

A comparison between S and R during a period, $t_1$ , before and a period, $_{t2}$ , after the changes in $L$ yields:

$$\bar{E}_S = (\bar{T}_{S,t2} - \bar{T}_{Rt2}) - (\bar{T}_{S,t1} - \bar{T}_{R,t1}) = \Delta\bar{T}_2 - \Delta\bar{T}_1 \qquad (2A)$$

Here it is assumed that the effects of macro-scale weather at *R and S are* nearly the same before and after the changes in $L$. It is also implied that at both sites the effects of the unchanged part of the local landscape are the same under similar weather conditions during different periods.

Effects of differences in general weather pattern during the periods before and after the changes in $L$ took place may be eliminated, or at least markedly reduced by grouping the data in classes e.g. according to cloud cover and wind, and/or temperature classes.

$\Delta\bar{T}_1$ and $\Delta\bar{T}_2$ and their standard deviations are computed for all groups, and the statistical significance of $\bar{E}_S$ is tested. An added requirement to the test is that the effect of changes in $L$ is considered statistically significant only when the sum of the standard deviations of $\Delta\bar{T}_1$ and $\Delta\bar{T}_2$ are smaller than $\bar{E}_S$ (Leith,1973).

## 6.2. Tables and estimation of energy fluxes

*Table A1. Saturation water vapour pressure at air temperature $T_a$ just above a water surface, $e^*_v$, and above an ice surface, $e^*_i$, and corresponding vapour content (absolute humidity), $\rho^*_v$ and $\rho^*_i$. $RH_i$ is relative humidity of saturated air above ice at temperature $T_a$.*

| $T_a$ | °C | -30 | -25 | -20 | -15 | -10 | -5 | 0 | 5 | 10 | 15 |
|---|---|---|---|---|---|---|---|---|---|---|---|
| $e^*_v$ | hPa | 0.51 | 0.81 | 1.25 | 1.91 | 2.86 | 4.21 | 6,11 | 8.72 | 12.27 | 17.04 |
| $e^*_i$ | " | 0.38 | 0.64 | 1.03 | 1.65 | 2.60 | 4.01 | 6.11 | | | |
| $\rho^*_v$ | $gm^{-3}$ | 0.45 | 0.70 | 1.07 | 1.60 | 2.36 | 3.41 | 4.85 | 6.81 | 9.40 | 12.83 |
| $\rho^*_i$ | " | 0.34 | 0.55 | 0.88 | 1.39 | 2.14 | 3.25 | 4.85 | | | |
| $RH_i$ | % | 75 | 79 | 82 | 85 | 91 | 95 | 100 | | | |

### 6.2.1. Energy exchange at a water surface

The main components in the energy exchange between the surface and the atmosphere are given by the equation:

$$(K\downarrow - K\uparrow) + (L\downarrow - L\uparrow) = K^* + L^* = Q^* = Q_G + Q_H + Q_E \qquad (3A)$$

where

$K\downarrow$ is short-wave, solar radiation to the surface (daytime)

$K\uparrow$ is reflected short-wave radiation

$K^* = K\downarrow - K\uparrow$ is net short-wave radiation to the surface

$L\downarrow$ is long-wave radiation from the atmosphere (day and night)

$L\uparrow$ is long-wave radiation from the surface (day and night)

$L^* = L\downarrow - L\uparrow$ is net long-wave radiation

$Q^*$ is net radiation (net gain or loss by radiation)

$Q_G$ is transfer of sensible heat to/from the sub-surface layer

$Q_H$ is turbulent transfer of sensible heat to/from the atmosphere

$Q_E$ turbulent transfer of latent heat (evaporation/condensation) to/from the atmosphere

Methods for estimating transfer of *latent heat* ***(evaporation)***, ***sensible heat***, and ***long-wave radiation*** are given below.

***Evaporation*** from a water surface is determined by the temperature difference between water and air, the humidity of the air, the wind speed, and the stability of the air above the water surface. In the empirical formula (after Ryan et al., 1974), that was applied to estimate the values given in Table A2 below, the stability is expressed by the term $N_F(T_v - T_a)^{1/3}$:

$$E = Q_E/L = 1/L[N_E u_2 + N_F(T_v - T_a)^{1/3}](e^*_v - e_a) \ mm\ s^{-1} \quad (4A)$$

where

$L$ is latent heat of vaporization (2.5 Joule $kg^{-1} \times 10^6$)

$u_2$ is wind speed 2 m above the surface (m $s^{-1}$)

$T_v$ is temperature of the water surface (°C)

$T_a$ is air temperature 2 m above the surface (°C)

$e^*_v$ is saturation vapour pressure at $T_v$ (hPa)

$e_a$ is actual vapour pressure 2 m above the surface (hPa)

$N_E$ is an empirical constant [3.24 W $m^{-2}$(hPa)$^{-1}$(m $s^{-1}$)$^{-1}$] (adjusted to wind measured 2 m above the surface)

$N_F$ is an empirical constant [2.7 W $m^{-2}$(hPa)$^{-1}$(°C)$^{-1/3}$]

The values obtained by means of (4A) multiplied by 86.400 (seconds per day) gives $E$ in mm $day^{-1}$.

As cold air moves over warmer water, the temperature and the actual vapour pressure of the air gradually increase, and accordingly the rate of evaporation decreases. Therefore, under these circumstances measurements at the up-wind edge a of wide waterbody applied in formula 4A will give too high evaporation values.

*Table A2. Diurnal evaporation (mm) from a waterbody with surface temperature $T_v$ = 1°C, calculated by formula (4A) for different air temperatures, $T_a$ , and wind speeds, $u_2$ , and for relative humidities, RH , 70 og 90% respectively.*

| $T_a$ °C | $u_2$ m $s^{-1}$ | 0 | | 2 | | 4 | | 6 | |
|---|---|---|---|---|---|---|---|---|---|
| | RH % | 70 | 90 | 70 | 90 | 70 | 90 | 70 | 90 |
| 0 | | 0.21 | 0.10 | 0.73 | 0.34 | 1.24 | 0.58 | 1.75 | 0.82 |
| - 5 | | 0.56 | 0.43 | 1.38 | 1.04 | 2.18 | 1.66 | 2.98 | 2.29 |
| -10 | | 0.94 | 0.83 | 1.96 | 1.71 | 2.98 | 2.61 | 3.99 | 3.50 |
| -15 | | 1.23 | 1.14 | 2.40 | 2.21 | 3.56 | 3.30 | 4.73 | 4.39 |
| -20 | | 1.46 | 1.40 | 2.74 | 2.64 | 4.01 | 3.84 | 5.29 | 5.05 |
| -25 | | 1.65 | 1.60 | 2.99 | 2.90 | 4.32 | 4.20 | 5.66 | 5.50 |

Several formulas for estimating the transfer of ***sensible heat***, $Q_H$, from a water surface to the air above have been developed. If the transfer of ***latent heat***, $Q_E$ , is known, $Q_H$ can be estimated by means of the ***Bowen-ratio*** , $\beta$ , (Andreas et al., 1979):

$$\beta = Q_H/Q_E = c_p\kappa_H(T_v - T_a)/L\kappa_E(q^*_v - q_a) \qquad (5A)$$

where $c_p$ = 1010 J $kg^{-1}$ $K^{-1}$ is spesific heat of air at constant pressure, $\kappa_H$ = 19.0 $m^2$ $s^{-1}$ x $10^{-6}$, and $\kappa_E$ = 21.3 $m^1$ $s^{-1}$ x $10^{-6}$ are molecular diffusion coeffisients of air for sensible and latent heat respectively at air temperature 1°C, $L$ = 2.5 J $kg^{-1}$ x $10^6$ is latent heat of vaporization, and $q^*_v$ is spesific humidity of saturated air at temperature $T_v$, and $q_a$ is actual spesific humidity of the air. In the Tables A2 og A3 a is referred to 2 m above the surface.

The evaporation values in Table A2 above multiplied by $\beta$ (Table A3), and by 29 W $m^{-2}$ gives estimated $Q_H$ values in W $m^{-2}$.

*Table A3. The Bowen ratio, (5A), above water with surface temperature $T_v$ = 1°C, at different air temperatures, $T_a$ , and relative humidity, RH, 70 0g 90% respectively.*

| RH % $T_a$°C | 0 | -5 | -10 | -15 | -20 | -25 |
|---|---|---|---|---|---|---|
| 70 | 0.25 | 0.94 | 1.37 | 1.73 | 2.10 | 2.46 |
| 90 | 0.54 | 1.23 | 1.57 | 1.87 | 2.19 | 2.53 |

The ***emission of long-wave radiation*** from a water surface is:

$$L\uparrow = \epsilon \sigma T_v^4 \quad W\ m^{-2} \qquad (6A)$$

where $\epsilon$ is emissivity (for water 0.97-0.99), $\sigma$ = 5.67 x $10^{-8}$ W $m^{-2}K^{-4}$ (Stefan-Boltzman constant), and $T_v$ is the absolute temperature of the surface. *If* $\epsilon$ = 0.97 and $T_v$ = 1 °C = 274.2 K, the emission, $L\uparrow$, is 320 W $m^{-2}$. The ***long-wave counter radiation,*** $L\downarrow$ , from the atmosphere by clear skies can be estimated e.g. by means of an empirical formula developed by Swinbank (1963):

$$L\downarrow = 5.31x10^{-13}T_a^6 \quad W\ m^{-2} \qquad (7A)$$

where $T_a$ is the air temperature [K]. By clear skies $L\uparrow$ is always markedly greater than $L\downarrow$ ; i.e. $L^* = L\downarrow - L\uparrow$ is negative, and the surface loses energy by long-wave radiation. Clouds increase the atmospheric counter radiation, and by overcast skies - especially with low clouds - the radiative loss is small.

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