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Ice thickness estimates for design of structures

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ABSTRACT

The level ice thickness is perhaps the most important ice property in both engineering and geophysics. Ice action from level ice scales almost linearly with ice thickness and the level ice thickness gives the mode in ice thickness distributions used in climate models. The consolidated layer thickness and probably also the keel depth in first-year ridges are functions of level ice thickness. In this paper we examine formulas based on Stefan's law to determine the maximum level ice thickness. Some of the underlying thermodynamics is explained and related to the different coefficients in the formulas. Traditionally the accumulated air temperatures (FDD) can be used, but then only an upper limit of ice thickness can be given. In addition one should know (or estimate) when the ice forms and starts to grow and how much snow that deposits on the ice. The equations given by \citet{iso:2019}, the Norwegian Road administration (N400), the Danish rules for the Baltic, the Russian SNIP and guidelines for inland waterways in USA are compared and discussed. The thinner the maximum level ice is, the more difficult is it to make good estimates. Finally, some additional complications are discussed.

KEY WORDS Ice thickness, Design of marine structures, Basic formulations; Standards

INTRODUCTION

The ice thickness goes into almost any formula or method to estimate ice action on fixed and floating structures (see e.g. ISO19906, 2019). The ice action, or force, usually scales almost linearly with ice thickness, making it a vital parameter to estimate. The level ice thickness is further a key parameter in any characterization of a floating ice cover and can be used to estimate the thickness of the consolidated layer and the keel depth of first-year ridges (Lepparanta et al., 1995; Høyland, 2002; Amundrud et al., 2004; Samardzija and Høyland, 2023). It is a complex and dynamic variable that is influenced by a wide range of factors, including temperature, precipitation, and oceanic flux. Rather than being a single, static value, ice thickness is better understood as a distribution of various thickness values that are constantly (but rather systematically) changing throughout the course of a season, and has a spatial variability. It is essential to specify the type of ice thickness under examination when conducting measurements, as well as when determining design ice thickness values for ice load calculations, to ensure accurate results.

There are spatial and temporal scales in ice thickness estimation. If the application is to estimate ice action on structures one would want to know the ice thickness on the scale of the structure (10 to 100 m) or a wind park (some km). However, since many structures are located in coastal areas, close to shipping channels or form a wind-mill park, the local boundary conditions become important. In some cases the presence of the structure (or structures) may change the ice conditions. If we move to temporal scales we should distinguish between ice thickness estimation for design or operation. In the latter case the relevant temporal scale is the duration of the operation (often hours to days) whereas in the design the ice condition over the life-time should be estimated, typically 50 years. If we narrow in further and consider only design of structures we still need to separate between fatigue and extreme static action (overall structural stability). In the first case one needs a distribution of all thicknesses (thickness and their duration) over the lifetime, whereas extreme static load requires an estimation of extreme ice thickness during lifetime. This paper is mostly about estimation of ice thickness for design of marine structures.

Ice actions estimation involve both uncertain parameters and uncertain models, and can be done with deterministic or probabilistic methods. Deterministic methods rely on the assumption of a single, specific value of ice thickness, typically that of the extreme maximum level with a defined return period. On the other hand, probabilistic methods take into account the inherent uncertainty of ice thickness by considering the distribution of values that incorporate both spatial and temporal variations. The inclusion of inter-annual and seasonal variations of ice thickness in probabilistic calculations provides a more robust and comprehensive assessment of ice loads, as it accounts for the variability of ice thickness in space and time.

When estimating the ice thickness the best is to use locally measured data to fit a function and then use this to evaluate trends in climate change. In this paper we will look at simple and physically-based equations that can be used to estimate ice thickness. The ice thickness and growth depend on a range of factors and a precise estimation is not easy, but we will argue that the maximum ice thickness can be expressed as a function of the accumulated freezing degree days in line with the classical paper Stefan (1891):

$$h_i^2 + a h_i = b\left(FDD - c\right) \qquad [\mathrm{m}^2] \tag{1}$$

where *a*, *b*, and *c*, are coefficients that have physical meaning and will be discussed in this paper, and *FDD* has the unit °C Days and is defined as:

$$FDD = -\sum_{days} \overline{T_a}\ , \qquad \overline{T}_a < 0°C \qquad [\,°C \cdot days\,] \tag{2}$$

where $\overline{T_a}$ is the daily average air temperature. Note that only negative air temperatures contribute, and positive air temperatures should be excluded. This strictly applies to fresh water, because sea water contains salt so that its freezing point is approximately -1.9 °C and the saline ice contains liquid brine. However, it seems to work well also for sea ice. With such a simple model the extreme vales for ice thickness simply follow the extreme values for air temperatures, and here relatively reliable and long-term data exists.

PHYSICAL MECHANISMS AND PROCESSES DURING FORMATION AND GROWTH OF LEVEL ICE

2.1. Onset of formation

Ice will form only when surface water temperature reaches the freezing point ($T_w = T_f$). This depends on several factors, but the water depth, tidal (or other) current, islands and river run off are key factors. The onset of ice formation is generally difficult to model and predict so that local knowledge and data is essential. In some coastal areas such as fjords, ice may form but then drifts out into warmer offshore waters by wind and currents (many fjords on the west coast of Spitsbergen). Islands act as anchor point for sea ice and increases the possibility for developing and ice cover (Aaland archipelago in the Baltic). In several Norwegian fjords the ice seems to form from almost fresh surface water originating from rivers that freeze before it is mixed down by wind and waves (O'Sadnick et al., 2022).

2.2. Ice growth on the bottom and oceanic flux

Once a thin initial layer of ice is formed it will start to grow downwards as long as the freeboard is positive and the heat transfer up into the atmosphere is larger than the oceanic flux coming up from the water. This process often produces columnar ice. Any heat transfer from the ocean or the water below (oceanic flux q_{ocean}) will reduce the ice thickness. The oceanic flux is a function of the temperature, velocity and salinity of the water and no simple formulas can be used to quantify it. Partly because of complicated physics, but also because of lack of good data. Since saline water is heavier than fresh water, warmer water may exists deep down, but not give a noticeable oceanic flux as the stable stratification prevents convection.

2.3. Snow and ice growth on the top

The snow depth usually increases during a winter season and is an important factor as it insulates the top ice surface from the atmosphere, reduces the temperature gradient in the ice so that the ice growth rate is reduced. However, the snow can have another important effect and that is to create negative freeboard so that part of the snow-pack is soaked in sea water. This water insulates the ice from the cold atmosphere completely, reduces the vertical temperature gradient through the ice cover and also the ice growth. Now ice forms from the top of the soaked snow and grows downwards towards the original ice cover. As long as the weather is cold enough this layer will freeze-up and the bottom ice formation continues. However, in many populated areas (such as Norwegian fjords) the soaked snow does not freeze up before the next precipitation event and several icy layers form. This happens for limited *FDDs* combined with substantial precipitation. This process creates layers of ice with different strength and the traditional concept of level ice thickness breaks down (Hornnes et al., 2023). Even in the Arctic basin top ice has been observed (Provost et al., 2016).

2.4. Seasonal trend and duration of an ice cover

Ice thickness has in most cases a clear seasonal trend, so that it grows, reaches a maximum and decreases. To develop an ice thickness distribution over many years it may be sufficient to assume the same ice growth function every year and tune it with the freeze-up time and FDD as random variables. In areas with infrequent ice occurrence such as the Southern Baltic the ice thickness distribution seems to have a smaller tail, and this probably has to do with different conditions for break-up. Sooner or later the ice starts decaying, often provoked by a combination of solar radiation and warm air and water. In some cases it breaks up and is transported into warmer and more open water where it melts, and in other enclosed areas it more or less melts in place.

SOME BASIC THERMODYNAMICS AND EQUATIONS

3.1. Atmospheric heat transfer and radiation

The energy exchange through the top surface consists of long - and shortwave radiation, convection and sublimation. The net surface heat flux (q_{atm}) can be written as (Maykut, 1986):

$$q_{atm} = q_{LW} + q_{SW} + q_{conv} + q_s + q_m \qquad [\mathrm{W}] \tag{3}$$

where q_{LW} is the net longwave radiation, q_{SW} is the net shortwave radiation, q_{conv} is the convection, q_s is the sublimation and q_m is the top surface melting heat flux. Application of Eq. 3 requires input data such as air temperature, wind, cloudiness, relative humidity and other parameters. Such data may be accessible from remote sensing even if weather station data is limited. However, the estimation of extreme ice thickness with so many relatively uncertain parameters becomes complicated and uncertain. A simpler model developed by (Adams et al., 1960) only needs the air temperature history and possibly wind expressed through some kind of convection model:

$$q_{atm} = H_{ia}\left(T_s - T_a\right) \qquad [\mathrm{W}] \tag{4}$$

where T_a and T_s are the air and the top ice-surface temperatures and H_{ia} an empirical parameter given (Adams et al., 1960):

$$H_{ia} = \max\left[11.6; 5.7 v_a^{0.8}\right] \qquad \left[W / {}^\circ C\right] \tag{5}$$

where v_a is the wind velocity at 10 m altitude. Salganik et al. (2021) compared Eqs.3 and 4 with field data and found a good match.

3.2. The base case, immediate freeze-up and ice surface temperature equal air temperature

Let us start with the assumptions behind Stefan's law (Stefan, 1891). These are no snow, no oceanic flux, ice surface temperature equal to air temperature, linear temperature gradient through the ice thickness, no radiation and water temperature at freezing point. Then it is strait forward to derive the following equation:

$$h_i^2 = \frac{2k_i}{\rho_i l_i} FDD \cdot \alpha \qquad [\mathrm{m}^2] \tag{6}$$

where h_i is the thickness, k_i, ρ_i, and l_i are respectively the thermal conductivity, the density and the latent heat of the ice, *FDD* accumulated freezing degree days (°C days) and α=86400 s/days to get the correct unit. In general Eq. 6 over-predicts the maximum ice thickness for two reasons. It assumes that:

- The surface water temperatures are at freezing point when air temperature drops below zero. In general, it is the air temperature that cools the water and this usually takes some time.

- The ice surface is as cold as the air temperature. In reality both snow cover and the convective flux between atmosphere and ice relaxes this assumption. Below we will look different ways of improving the precision of prediction.

3.3. Constant snow thickness

Let us now assume a constant snow thickness h_s. The snow cover increases ice surface temperature and improves prediction considerably. In this case we keep *b* and *c* defined in Eq. 1 as above in Eq. 6, but *a* becomes no-zero:

$$h_i^2 + \frac{2h_s k_i}{k_s} h_i = \frac{2k_i}{\rho_i l_i} FDD \cdot \alpha \qquad [\mathrm{m}^2] \tag{7}$$

where h_s is the snow thickness, k_s is the thermal conductivity of the snow and the other parameters are as in Eq. 6. Inclusion of realistic snow thickness over the season improves predictions considerably, but it is difficult to quantify the snow depth.

3.4. Convection in the air

The other option of using more realistic ice surface temperature is to introduce convective heat transfer between the air and the ice and return to assuming no snow (Fig. 1). Then we keep b and c as above but find:

$$h_i^2 + \frac{2k_i}{H_{ia}} h_i = \frac{2k_i}{\rho_i l_i} FDD \cdot \alpha \qquad [\mathrm{m}^2] \tag{8}$$

where the empirical coefficient H_{ia} is given above. We should note that for small values of hi the linear term dominates so that the ice growth becomes liner with *FDD*. When the ice reaches a few decimeters it shifts to square root dependency.

3.5. Oceanic flux and counting FDDs from freeze-up

Any oceanic flux will delay ice formation and reduce ongoing ice growth rate. The delayed freeze-up is in many cases the vital effect and mathematically it very simple to adjust for this. One simply subtracts the *FDDs* applied to cool down the water ($FFDc = FDD - c$) and start counting from freeze-up. Note that in deep saline waters it may take long time to cool down all the water and in many places hardly any ice forms at all (Trondheim fjord). In other coastal areas where there is river run out, cold water may come from mountains (where air temperatures are lower) and create stratification so that only a few local negative *FDDs* may be required to form ice. Local information on how often and how long the ice cover exists will be very useful. Any oceanic flux during ice growth will of course reduce the ice growth rate and may be included mathematically in a the same way as the convection in the air. It is not so easy to quantify and we think it is less important that quantifying a realistic freeze-up date and precipitation. Finally, we should also note that supercooled sea water may lead to formation of platelet ice and as such increase the ice thickness (Katlein et al., 2020).

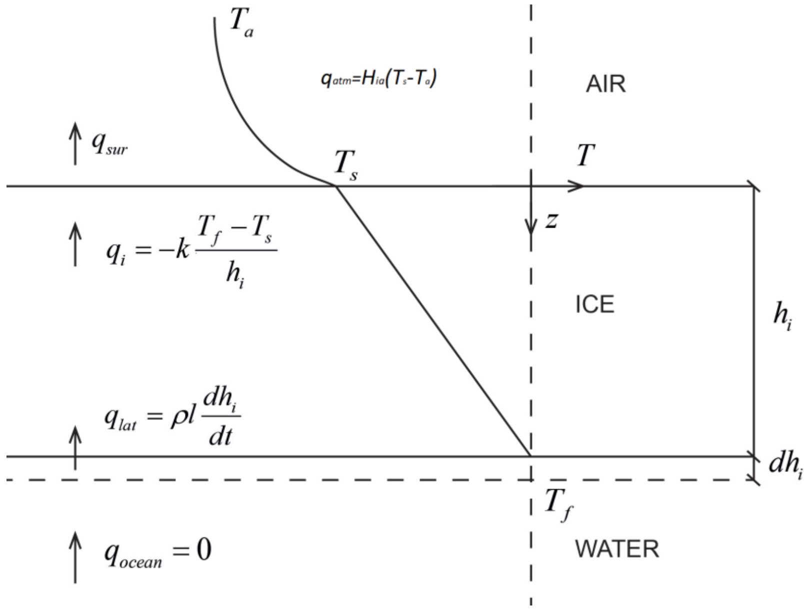

Figure 1. Illustration of Stefan's law with convection in the air.

EXAMPLES FROM GUIDELINES

There are numerous papers expressing empirical formulas and calibrating expressions to measured data and this paper does not provide an overview of these. Here we only present formulas that we have found in different standards.

4.1 Standard approach

An easy way is to lump all the missing physical conditions into one empirical factor (sometime called ω), multiply it with the right-hand side of the Eq. 6 and apply Eq. 9. With this approach one needs only predictions of *FDD*.

$$h_i^2=\omega\frac{2k_i}{\rho_i l_i}FDD\cdot\alpha \qquad [\mathrm{m}^2] \qquad (9)$$

where ω is between 0 and 1 and must be derived from measured data. In general, the more snow the lower ω. As long as the data are measured under similar conditions as the ones on the site of interest this will give good predictability. However, if the precipitation, water depth and other conditions are different the results will not be good. The local oceanographic flux is also difficult to determine, but for many shallow areas it is small. In Eq. 9 it is not taken into account directly, but lumped into ω.

4.2 Russian SNIP

Zubov's expression (Zubov, 1943) is purely empirical but much used:

$$h_i^2 + 50h_i = 8FDD \qquad [\mathrm{cm}^2] \tag{10}$$

note that h_i is given in [cm] so that a and b in an equation that gives ice thickness in meters become respectively 0,5 and $8 \cdot 10^{-4}$. The main data Zubov used to calibrate his model were collected at Uyedineniye Island in 1935-1936 and at Cape Schmidt the winter of 1936-37 and *by using available observations made by certain other polar stations as a control* (Zubov, 1943). We may note that the formula corresponds to a constant average snow thickness of less than 0.1 m. This is of course very small, but as the thin ice grows faster and often without snow it cannot be used to compare with snow depth data directly. If we assume a total *FDD* of 3000 and assume that the first 1000 occurred without snow we need to have in average 30 cm of snow for the rest of the season to fit with Zubov (1943).

Another fit to data (Eq. 11) provided by Lebedev (1938) has also been used. This equation is not dimensionally correct and cannot be transferred into our standard form (Eq. 1).

$$h_i = 1.33 \cdot FDD^{0.58} \qquad [\mathrm{m}] \tag{11}$$

4.3 Danish waters in the Baltic

The following formula Tryde (1983) has been calibrated and validated with data from Danish waters, including the cold winters of 1939 - 1942:

$$h_i^2 = 0.03(FDD - 50) \qquad [\mathrm{m}^2] \tag{12}$$

The factor 0.03 corresponds to an ω of about 0.75. This is a quite high factor and would correspond to little snow and corresponds well with local conditions if we assume that ice forms in Danish water in quiet, dry and cold periods so that the ice growth is not limited by precipitation. The *FFDs* in Danish water has been less than 300ºCdays since 1960 but was 500ºCdays in 1941-42.

4.3 Norwegian Road administration

The rules of the Norwegian Road Administration (SVV) for ice action on bridges across fjords, lakes, and rivers suggest to estimate design ice thickness as:

$$h_i = \frac{1}{175}\sqrt{FDH} = \frac{1}{175}\sqrt{24FDD} \qquad [\mathrm{m}] \tag{13}$$

where *FDH* is Freezing Degree hours. It corresponds an ω of about 0.66 which gives similar thickness as Stefan's law with air convection (Eq. 8) for FDD up to about 1000ºCdays (if the average wind velocity is around 5 m/s). The formula is meant to apply for all Norwegian waters, that is both river, fjords and lakes. The conditions in different rivers and fjords in

Norway vary considerably and the use of a single ice thickness equation must give considerable over-estimations in some cases.

DISCUSSION

5.1 Guidelines

Figures 2 - 4 show the different formulas plotted for high, intermediate and small number of *FDDs* and includes some measured data. The first thing to notice is that there are substantial differences between the different formulas, even if we disregard the Stefan standard (Eq. 6) that clearly overestimates real ice thickness. Secondly, the relative differences are bigger for smaller *FDD*s, in other words it may be more difficult to give good predictions in lower latitudes. The conditions, especially the snow thickness, usually changes during the winter so an equation with constant parameters (a, b and c) should not fit throughout a long season, or for cases with quite different *FDD*s. Some experimental data are also shown in Figure 4; one relatively mild winter from Bothnian Bay, six Svalbard years and the high Arctic MOSAiC data. Five out of six Svalbard years fit nicely around Zubov's equation. The 2003/2004 season was different with a cold and dry December- January, and a warm March. In other words the ice growth took place largely without snow.

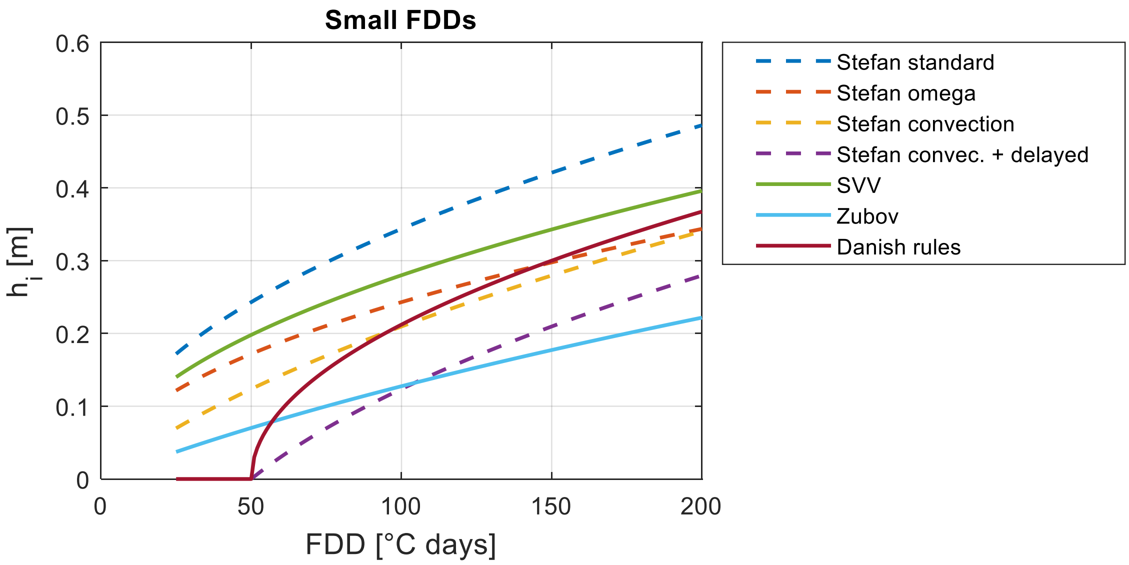

Figure 2. Different formulas for small *FDDs*.

Zubov's equation also fits quite well for the Bothnian Bay (Li et al., 2016). It is interesting to note that Zubov's equation overestimates MOSAiC data even though it was calibrated to similar *FDDs*. One could imagine that ice starts growing earlier on the shallow Russian shelf so that even for same *FDDs* the open Arctic basin should grow thinner ice. However, it may also be that there was more precipitation in 2019/2020 than in the late 30's, so that Zubov's calibration is outdated? The Danish rule has been calibrated to the icy winters in Danish Baltic waters and should fit well for the southern Baltic. We note that further east in the Southern Baltic the coastal waters are less confined by land, and one may defend using a somewhat lower coefficient *b* (Jørgensen et al., 2023). However, this equation should not be used for large *FDDs* as it overestimates the MOSAiC measurements by about 30%.

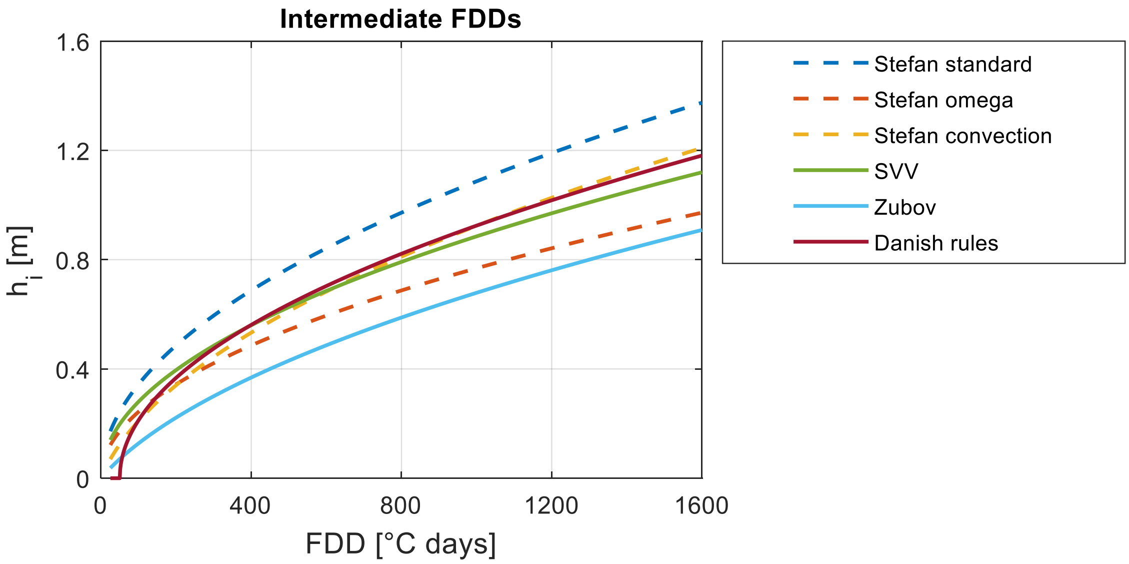

Figure 3. Different formulas for small *FDDs*.

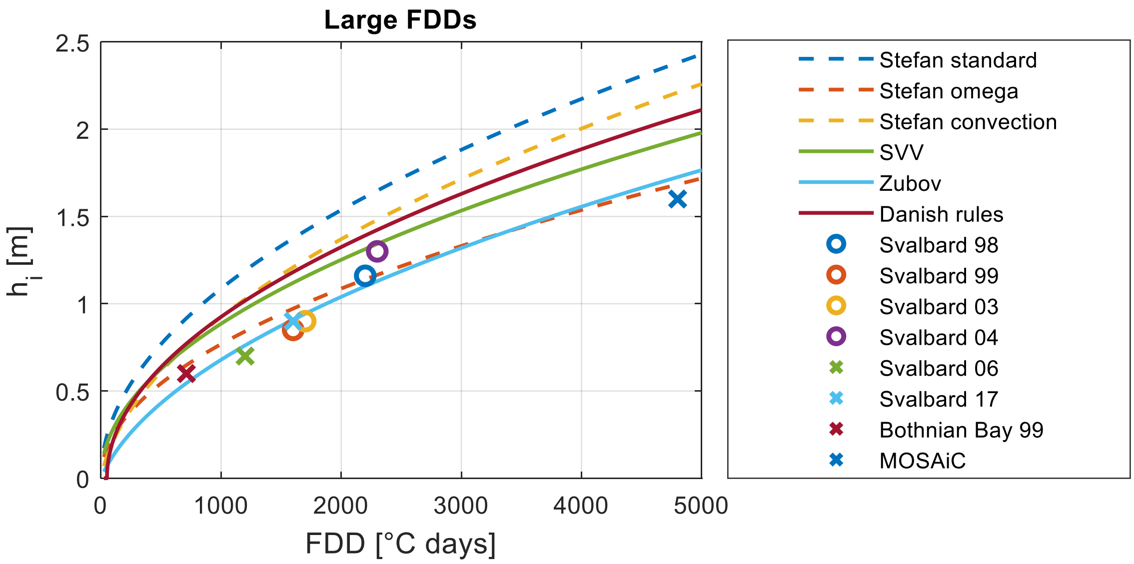

Figure 4. Different formulas for large *FDDs* and some empirical data.

5.2 Climate change

In general, the climate is changing so that it in average gets warmer and the ice becomes thinner (Stroeve and Notz, 2015, Lindsay and Schweiger, 2015, Holland et al., 2010). However, it is not clear what happens with the extremes, so care should be taken in prolonging average warming trends. Another important factor of climate change is that the existing formulas are calibrated with historical data and only works for predictions as long as the met-ocean parameters are approximately the same. Especially the relationship between precipitation and *FDD* may and making the existing parametrization inaccurate. Also, if the water becomes warmer, (such as the North Atlantic coastal current) the freeze-up may come later and ice gets thinner.

5.3 Complicating phenomena

Let us in the end look at some complicating factors that are not covered in this paper. Firstly, what happens when there is sufficient precipitation and limited FDDs as in many Norwegian (including Svalbard) fjords, the ice-snow cover may consist of layers of ice, slush and snow. In these conditions a single ice thickness is not easy to define (Hornnes et al., 2023).

Secondly, the ice thickness can increase due to ice deformation where both thicker ice and more open water is created. In the open water ice growth accelerates, but still creates a thinner level ice. The rafted and ridged ice is often called mechanically created ice thickness and can exceed what is predicted by formulas given in this paper. Once again local data and knowledge of local conditions are crucial to make reasonable and safe predictions of ice thicknesses. A simple formula for maximum thickness of the consolidated layer (h_c) in first-year ridge is (Leppäranta et al., 1995)

$$h_c^2 = \frac{h_i^2}{\eta_M} \quad [\mathrm{m}^2] \tag{14}$$

where η_M is the macro-porosity of the unconsolidated part of the ridge keel (ice rubble) and often range between 0.25 and 0.35.

Thirdly, the ice can drift long distances before reaching site of interest such as the southern Barents Sea or the Canadian East coast. The ice growth may happen under significantly more severe conditions than on site of interest, so that using local data will underestimate the ice thickness. In these cases, the ice thickness estimation becomes more challenging and has not been addressed in this paper. We can only point to that one needs to estimate ice thickness in areas where one believes that ice originates and grows, and then estimate the probability that this ice can drift and reach the site of interest. In many cases the melt-rate from oceanic heat becomes the key issue.

CONCLUSIONS

Physical equations based on Stefan's law and corresponding parametrization in different standards has been presented and discussed. The accumulated Freezing Degree Days (FDD) is the only geophysical input, but the physical assumptions behind the different constant are explained. The main conclusions are:

- The two major uncertainties in application of Stefan-type formulas are the freeze-up time and amount of snow.
- Local knowledge and data is important, and coefficients derived from sites with similar met-ocean conditions should be used.
- There are significant differences between the different equations, and the relative differences are largest for small *FDD*s.

- The Danish rules should be used in southern Baltic and Zubov's equation seems to give a good fit to data from Svalbard and the Bothnian Bay.

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