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The physics of ice sheets

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Abstract

The great ice sheets in Antarctica and Greenland are vast deposits of frozen freshwater that contain enough to raise sea level by approximately 70 m if they were to completely melt. Because of the potentially catastrophic impact that ice sheets can have, it is important that we understand how ice sheets have responded to past climate changes and will respond to present climate changes, so that we can predict the effect global warming will have on sea level rise in the coming centuries. The purpose of this article is to introduce students to some of the basic concepts of glaciology, the physical variables that control the evolution of ice sheets, and how changes in these parameters may affect the long-term evolution of ice sheets.

Overview

The glacier was God's great plough set at work ages ago to grind, furrow, and knead over, as it were, the surface of the earth.

-Louis Agassiz

Glancing through the research interests of any physics department you would get the impression that all interesting topics of current physics research are rooted in modern physics, i.e., quantum mechanics and general relativity. In contrast, the physics that governs the formation and dynamics of ice sheets and glaciers is fully grounded in classical, eighteenth century, physics with little reference to the more exotic brands of physics practised in most physics departments. For this reason it would seem that ice sheets should be fairly simple to understand using elementary physics. But our understanding of the mechanics of ice sheets is still quite poorly developed, with many outstanding questions that remain to be solved. The goal of this article is to present a short introduction to ice sheets, first qualitatively describing the great ice sheets in Greenland and Antarctica, and then showing how temperature and elevation control the formation and disintegration of ice sheets. We end with a discussion of some of the factors that control the velocity at which ice sheets flow, along with some of the topics of current glaciological research. Our focus here is introducing some of the physical processes at work. Interested readers should also see [4] for an excellent collection of photographs of glaciers, illustrating many of the ideas presented in this article.

The Greenland and Antarctic ice sheets

Ice sheets are large, thick frozen masses of ice that extend over an area greater than 50 000 km^2. The only ice sheets that remain from the previous ice age exist in Antarctica and Greenland, where they extend over most of the landmass. However, during previous ice ages, ice sheets also covered large areas of North America and northwest Europe, with the North American ice sheet extending as far south as Chicago.

The great ice sheets in Greenland and Antarctica contain nearly all of the world's ice and approximately 70% of the world's fresh water, locked up in frozen deposits that have accumulated over hundreds of thousands of years [7]. These

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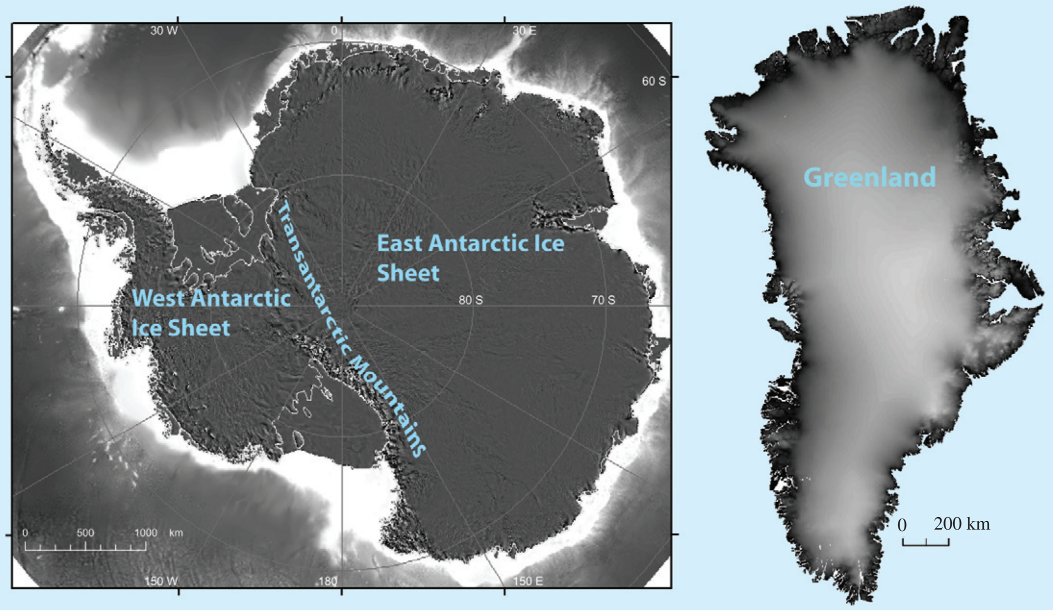

Figure 1. Left panel: MODIS image of Antarctica (provided courtesy of NSIDC [12]) showing the East and West Antarctic ice sheets. The two ice sheets are loosely separated by the Transantarctic Mountains. Right panel: Shaded relief map of the Greenland ice sheet. Light colours indicate higher elevations (DEM from [1]).

ice sheets contain an amount of frozen water that, if released into the ocean, would raise global sea levels by close to 70 m [1, 7, 8, 10]. Ice sheets are enormously effective at grinding down surface relief, causing new landforms to be carved out of the rock. For instance, the North American ice sheet excavated the great lakes as it retreated while, on a more subtle scale, the rich sandy moraines of Scotland were formed in the wake of the retreat of the European ice sheet.

The Antarctic ice sheet is composed of two parts, separated by the Transantarctic Mountains: the East Antarctic ice sheet and the West Antarctic ice sheet (see figure 1, left panel). Combined, the two ice sheets have an area of roughly 14 million km^2, close to one and a half times the area of the United States of America. The East Antarctic ice sheet is approximately 25 million years old and has an average ice thickness greater than 2 km. If the East Antarctic ice sheet were to melt completely, there is enough ice to raise sea level by more than 50 m. In contrast, the West Antarctic ice sheet is much smaller, with only enough ice to raise sea level by about 5 m. Despite its smaller size, most of the bedrock upon which the West Antarctic ice sheet sits is grounded beneath sea level [7, 10]. This means that it may be gravitationally unstable and much more sensitive to climatic perturbations than the larger East Antarctic ice sheet. Some evidence of the greater sensitivity of the West Antarctic ice sheet is that it has already lost two thirds of its volume since the last glacial maximum.

The Greenland ice sheet (figure 1, right panel) has an average thickness greater than 2 km, with enough ice to raise sea level by more than 7 m. Unlike the West Antarctic ice sheet, the bed of the Greenland ice sheet is predominantly raised above sea level. However, because the surface temperatures over most of the Greenland ice sheet are warmer than those over the Antarctic ice sheets, studies indicate that small changes in temperature, perhaps as small as 3 °C, may trigger irreversible melting of the Greenland ice sheet [2, 9].

Ice sheets, snow fall and the big squeeze

Ice sheets get thicker and expand as layers of snow are deposited in the interior of the ice sheet where it is coldest. This snow accumulates over millions of years. As the snow settles, more recent snowfalls bury it. As more snow accumulates above, the pressure increases and the snow becomes compressed into layers of ice. These layers of ice eventually start to deform and flow, like molten metal or pancake batter spreading out on a frying pan. The pressure of the ice in the interior forces the ice to flow outwards, towards the coast, at speeds of the order of metres per year ($\sim 1 \times 10^{-8}$ m s^{-1}). The slow-moving ice gets squeezed towards the coastline, through faster-flowing ice streams (rivers of ice of fast-flowing ice within the ice) and outlet glaciers where speeds can exceed hundreds of metres per year. Once the ice reaches the ocean, it can even spread out on top of the water, forming freely floating platforms of ice such as ice shelves and ice tongues from which icebergs detach.

During ice ages, when global temperatures decrease, the volume of ice sheets grows. Because the amount of water in the Earth/climate system is approximately constant, this means that sea level decreases. During warmer times, the volume of ice sheets decreases and sea level rises again. The potential to rapidly elevate sea level by melting the Antarctic and Greenland ice sheets provides one of the primary reasons for studying ice sheets and how they respond to climate change.

Temperature, accumulation and the snowline

It is probably obvious that a necessary condition to form an ice sheet is that some of the snow that accumulates during the winter must remain after the summer, at least on average over many cycles of snow accumulation and ablation. (Ablation is the term used by glaciologists to describe the snow that melts during the summer.) Therefore two variables that determine whether an ice sheet can form are (1) temperature and (2) snowfall. These two variables are not really independent—they are connected through the land–ocean–atmosphere climate system—but for our purposes we will treat them as independent. Let us start by thinking about temperature. The average solar radiation from the sun decreases towards the poles, and therefore the temperature also decreases towards the poles. This favours the formation of ice sheets at high latitudes where temperatures are more likely to be sub-freezing for a larger portion of the year. However, glaciers also form in tropical regions. This is because the temperature also *decreases* with increasing altitude. This decrease in temperature with elevation is called the lapse rate. The lapse rate will change somewhat with moisture content in the air, atmospheric dynamics, etc, but a good rule of thumb is that it decreases by approximately 10 °C for each kilometre of elevation gained—see figure 2(a).

Glaciers that form in the tropics form on top of mountains, where the altitude is high enough that temperatures are below freezing for enough of the year so snow/ice can accumulate. Applying this to the Greenland and Antarctic ice sheets, we deduce that the portions least likely to be affected by changing climates are the places at highest elevation that are closest to the pole. Summarizing, the two factors that control whether an ice sheet (or glacier) will form are the precipitation and temperature. But the temperature depends on *both* latitude and altitude. With this in mind it is convenient to define the snowline as *the boundary separating the region where all the snow that falls in the winter melts during the summer from the region where all of the snow does not melt each summer*. Figure 2(b) shows a cartoon illustrating the effect of latitude and altitude on the snowline. You can see that mountains protrude out over the snowline, enabling glaciers to form in tropical regions, and also that the snowline bends downwards towards the colder polar regions. Above the snowline (called the accumulation zone) ice accumulates over time. Beneath the snowline (called the ablation zone), ice is lost over time.

There are two things to point out about the connection between the snowline and accumulation. First, since ice accumulates above the snowline this causes the altitude of the ice to increase over time, thus further raising the surface of the ice sheet above the snowline. Second, the incessant accumulation of ice above the snowline cannot be maintained indefinitely. Eventually the ice will start to ‘spill’ out and flow downhill towards the ablation zone. Figure 3 shows a schematic of an initially ice-free region with a bed that slopes downwards gently. A portion of the ground is above the snowline, and therefore ice begins to accumulate in this region (figure 3(a)). As the snow continues to accumulate, the ice sheet becomes thicker. Eventually, the ice will start to flow downhill towards the ablation zone (figure 3(b)). To put this more precisely, denote the rate at which ice is lost due to ablation by $b(x)$ and the rate

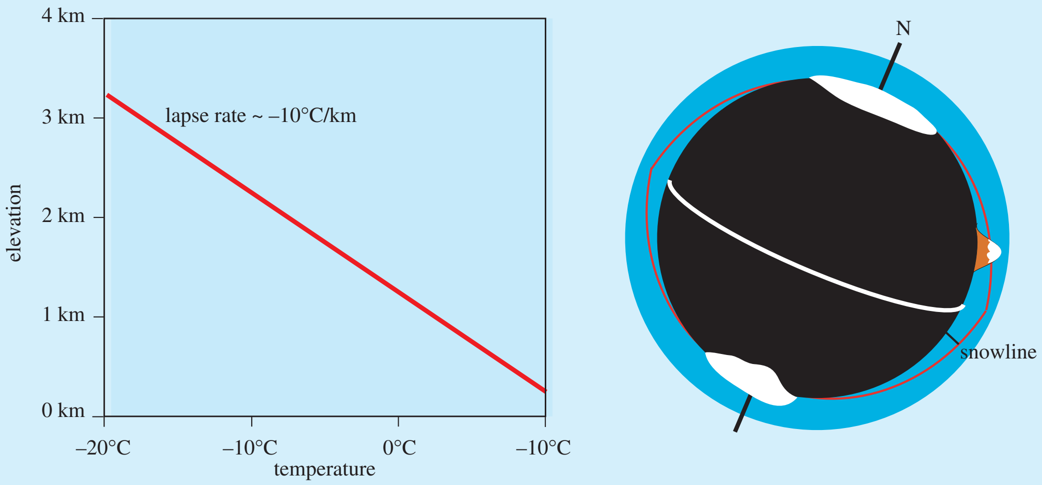

Figure 2. (*a*) Schematic showing how temperature varies with elevation. For this (fictional) situation, at an elevation of about 1 km the temperature is below freezing and ice sheets and glaciers can form. Below about 1 km the temperature is above freezing, making it impossible for a glacier or ice sheet to initiate at these low elevations. (*b*) Cartoon illustrating the effect of elevation and latitude on the snowline. Because the temperature of the atmosphere (usually) decreases with elevation, it is possible to form glaciers and ice sheets anywhere, even at tropical latitudes, if the glacier initiates at a high enough latitude, e.g. a mountaintop.

at which ice is gained by accumulation by $a(x)$ (note that accumulation and ablation have units of m s^{-1}). The snowline is then the point x_s at which $a(x) = b(x)$. Although $b(x)$ is greater than $a(x)$ in the ablation zone (by definition), there is a net flux of ice from the accumulation zone to the ablation zone: $q_s = h(x_s)u(x_s)$, where $h(x)$ is the ice thickness and $u(x)$ is the rate of movement of the ice. Therefore even though this region is below the snowline, there is still a net flux of ice into the ablation zone from the accumulation zone *even if the annual temperature in this region is above freezing*. The flow of ice (ice dynamics) prevents ice from accumulating indefinitely in the accumulation zone and moves it to the ablation zone, where it can melt. (Think of adding grains of sand to a sand pile. After the sand pile gets steep enough, sand grains start to fall down the slope.) It is important to realize that the snowline does not indicate the boundary separating regions where there is not any ice from regions where there is ice; it indicates the point where accumulation equals ablation.

How do ice sheets respond to climate change?

Now let us perform a thought experiment. We consider the same schematic situation as in figures 3(a)–(c). Now imagine that we move the snowline down (figure 3(d)). This can be accomplished by either (1) decreasing the temperature (and thereby decreasing $b(x)$) or (2) increasing the amount of snowfall (and decreasing $a(x)$). This increases the size of the accumulation zone and decreases the size of the ablation zone, thereby causing the ice sheet to advance, just as we would expect. In contrast, if the snowline moves up the reverse will occur (figure 3(e)), i.e., the accumulation zone will shrink, the ablation zone will grow, and the ice sheet will retreat. Finally, consider a third situation in which we increase temperature ($b(x)$) and accumulation ($a(x)$). In this case, determining whether the ice sheet advances or retreats depends on whether the increase in accumulation is larger or smaller than the increase in ablation. This implies that an ice sheet (or glacier) may grow in a warming climate if snowfall increases more than ablation (and vice versa). Moreover, an increased temperature will tend to increase the ablation most near the lowest-lying edges, where the temperatures are already warmest, causing the ice sheet to first erode around the edges. At the same time, a small increase in temperature in the interior of the ice sheet where the temperatures are substantially below zero may have very little effect

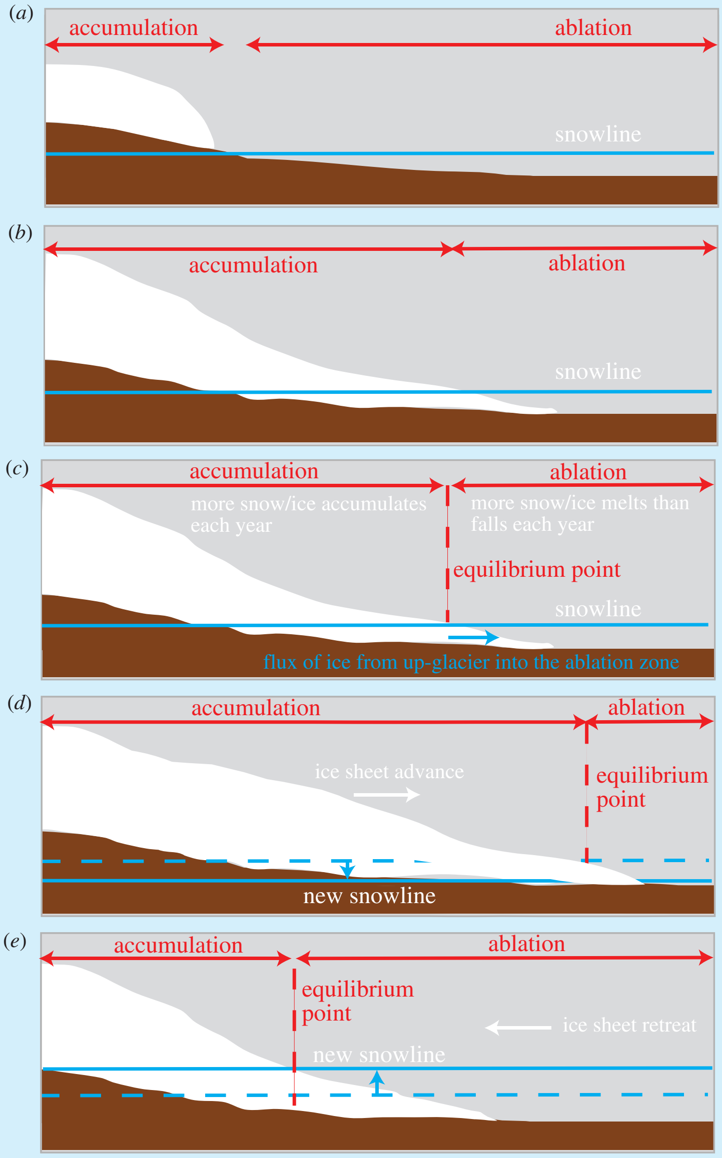

Figure 3. Schematic showing how an ice sheet forms. (*a*) Initially, the region above the snowline begins to accumulate snow and ice. (*b*) As more ice accumulates, the ice gets thicker and starts to flow outwards into the ablation zone. (*c*) Finally, a steady-state is reached. Panels (*d*) and (*e*) show the effect of moving the snowline down (*d*) and moving the snowline up (*e*). The blue dashed line indicates the old snowline while the solid blue line indicates the new snowline.

on the mass balance of the ice sheet. These cold, dry regions tend to be dominated by changes in accumulation. There is nothing inconsistent about an ice sheet that is simultaneously thickening in the interior and thinning along the lower-lying edges. This is a point worth emphasizing, because it has significant bearing on what is occurring today in Greenland, where thinning is occurring along the edges, due probably to increasing temperatures, while the interior of the ice sheet is relatively stable, and may even be thickening slightly [3, 5, 6, 11].

Mass balance and ice sheet dynamics

Until now we have considered the response of ice sheets to changes in temperature and accumulation, acknowledging that ice sheets flow, but ignoring the details of how such deformation occurs. This can also be an important factor in determining the mass balance of ice sheets. Let us denote the difference between accumulation and ablation by $m(x)$. Ultimately, an increase (decrease) in the integral of $m(x)$ over the ice sheet results in an increase (decrease) in the total volume of the ice sheet. However, ice dynamics, i.e., changes in the velocity u, may also be important in determining the shape of the ice sheet. To examine this further let us start by considering some factors that contribute to shifting the equilibrium line. Recall that the flux of ice across the equilibrium line is given by the product of ice thickness and (depth-averaged) velocity:

$$q_\mathrm{s} = h(x_\mathrm{s})u(x_\mathrm{s}). \tag{1}$$

In the steady state, this must be balanced by the integral of $m(x)$ from the head of the ice sheet at $x = x_0$ to x_s:

$$q_\mathrm{s} = h(x_\mathrm{s})u(x_\mathrm{s}) = \int_{x_0}^{x_\mathrm{s}} m(x)\,\mathrm{d}x. \tag{2}$$

We can rearrange equation (2) to find an equation for the steady-state ice thickness at $x = x_\mathrm{s}$:

$$h(x_\mathrm{s}) = \frac{1}{u(x_\mathrm{s})}\int_{x_0}^{x_\mathrm{s}} m(x)\,\mathrm{d}x. \tag{3}$$

We see that an increase in ice flow velocity will also cause the ice sheet thickness to decrease. This may cause the ice sheet to expand, but maintain the same total volume or, depending on how $m(x)$ changes with the change in ice thickness, the ice sheet may increase or decrease in volume. The lesson here is that finding that the ice sheet is thickening (or thinning) in one place is not enough to tell us about the mass balance of the entire ice sheet; we need to consider the integral over the entire ice sheet to separate the changes in ice dynamics from changes in the mass balance.

Ice dynamics and the role of meltwater

The next question that we wish to address is: what causes the ice to flow? This is a question that requires some knowledge of fluid mechanics to pursue. Unfortunately most physics students, even those at advanced stages of undergraduate study, are unfamiliar with stress, strain and tensor manipulation. We therefore adopt a somewhat simplified approach based on gravitational potential energy and a rigid block sliding down an inclined plane. Recall that the force due to gravity can be expressed as the gradient of gravitational potential energy:

$$F = -\nabla U. \tag{4}$$

The potential energy density (i.e., potential energy per unit volume) of a blob of ice of density ρ at height z above the bed is then

$$\frac{U}{\text{Volume}} = \rho g z. \tag{5}$$

Integrating the potential energy density over the ice thickness, $h(x)$, we find

$$\Theta = \frac{U}{\text{Area}} = \int_0^h \rho g z\,\mathrm{d}z = \rho g\frac{h^2}{2}. \tag{6}$$

This represents the gravitational potential energy of a column of ice of thickness $h(x)$. Let us define a quantity τ_d that has units Force/Area, which in analogy with equation (4) is defined by

$$\tau_d = -\nabla\Theta. \tag{7}$$

Taking the gradient of (6), we find

$$\tau_d = -\frac{1}{2}\rho g\frac{\partial h^2}{\partial x} = -\rho g h\frac{\partial h}{\partial x}. \tag{8}$$

This quantity, which is called the driving stress, is the continuum analogy of the force due to gravity. It increases with both the ice thickness and ice thickness gradient. As a simple model let us assume that the driving stress is entirely balanced by friction at the bed, neglecting the resistance due to viscous stresses within the ice. The exact form to use for the friction caused by the ice sliding over its bed is complicated and controversial, but we can gain some insight into the behaviour of ice sheets using a simple law of the form

$$\tau_f = \beta^2 u^p, \tag{9}$$

where β^2 is the coefficient of friction, u is the velocity, and p is a positive empirically derived exponent [7, 10]. Equating equations (8) and (9) and solving for the velocity, we find

$$u = -\left[\frac{\rho g h}{\beta^2}\frac{\partial h}{\partial x}\right]^{1/p}. \quad (10)$$

In words, equation (10) says that the stress or force per unit area that causes ice to flow is proportional to the ice thickness and slope of the ice. When combined with the friction law, we see that thicker ice and a larger thickness gradient cause the ice to flow more quickly. Equation (10), in a nutshell, explains why ice flows from the accumulation region (where it is growing into a large blob of ice of increasing thickness) into the ablation region.

Now let us consider the effect of lubricating the base of the ice by allowing meltwater from the surface to percolate down to the base. If the base is well lubricated we expect that the friction will be very low. Let us adopt a modified friction law of the form

$$\tau_f = \beta^2(\rho g h - P_{\mathrm{w}})u^p, \quad (11)$$

where P_{w} is the pressure of water at the base of the ice. Again, solving for the velocity, we find

$$u = -\left[\frac{\rho g h}{\beta^2(\rho g h - P_{\mathrm{w}})}\frac{\partial h}{\partial x}\right]^{1/p}. \quad (12)$$

As P_{w} increases, the velocity increases. This analysis breaks down as the ice approaches flotation, where $P_{\mathrm{w}} = \rho g h$, but it illustrates the important role that water at the base of the ice sheet can play in ice dynamics. See [8] for additional discussion on the role of how water at the base of ice sheets, in particular how water beneath several kilometres of ice, can be detected.

Conclusions

Ice sheets respond to changes in both temperature and precipitation. This can lead to unintuitive—but easily explained—behaviour, such as a warming planet that causes an ice sheet (or glacier) to thicken in the centre while simultaneously thinning along the margins. Temperature and precipitation are not, however, the whole story. Although ice may flow very slowly, ice sheets are dynamic, and changes in the velocity may also contribute to changes in the mass balance of the ice sheets. One mechanism by which this can occur is if meltwater penetrates to the base of the ice sheet, thereby lubricating the base and allowing the ice sheet to flow more quickly. Ice sheets are considerably more complicated than indicated by our analysis, with many outstanding research questions remaining. For instance, we still do not understand how pools of surface meltwater penetrate to the base of the ice sheet, although meltwater certainly does (see [4] for some spectacular photographs). Despite these uncertainties, considerable progress has been made over the past decade in understanding the physics that governs these processes [11], and the physics that we use to understand ice sheets is based entirely on good old-fashioned eighteenth century classical mechanics and thermodynamics.

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