REMOTE SENSING FOR NATURAL HAZARD STUDIES

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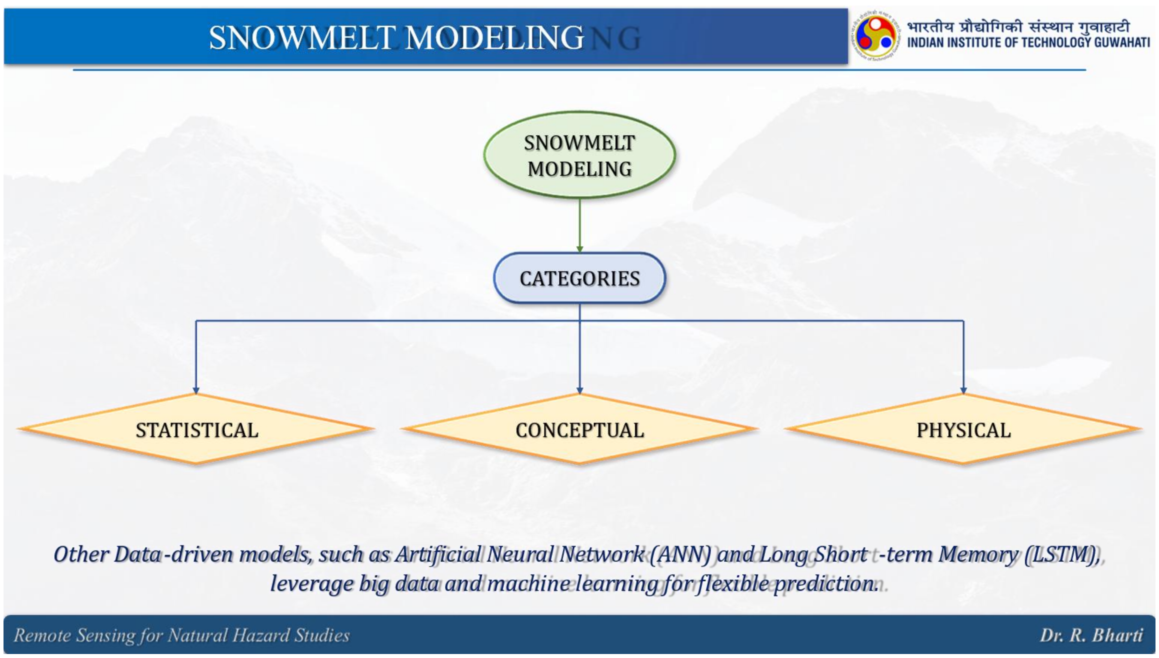
Lec 21b: Dynamics of Snow and Glacier Part B

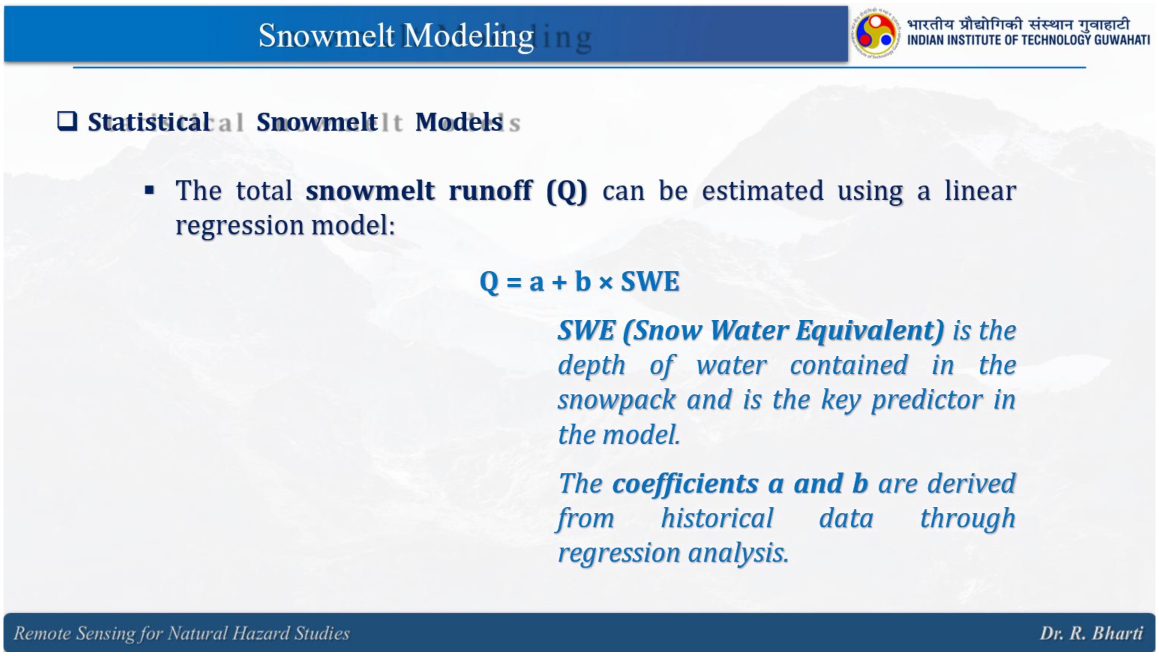
Hello everyone, welcome back to Lecture 21. So, we will continue this discussion. So, this is part 2. So, as we know, we were talking about the glacial dynamics. Snow melt modeling is another aspect of glacial dynamics that is very important. Snowmelt models simulate snow accumulation, melting, and runoff processes.



These processes are complex in nature because energy and water exchange occur between snow, soil, and the atmosphere. So, there is an exchange of energy. So, because of this, this modeling is very complicated. Snowmelt modeling is vital for predicting floods, especially in mountainous regions, because changes in temperature or any other climatic conditions cause glaciers to react, and due to the slopes of these glaciers, which are located at high altitudes, they can have a significant impact.

And, if there is a change in the climatic conditions, they respond, and because of that, there will be more melting, which will subsequently affect the low-lying areas. If the melting rate is high, then the low-lying areas will be in trouble, and we will have a flood-like situation. Accurate simulation helps to reduce flood risk and protect infrastructure and communities. Accurate simulation helps reduce flood risk and protect infrastructure and communities. So, this snow melt modeling can be categorized into 3 major parts.

The first one is statistical, then we have conceptual, and then we have physical. We will see them one by one. So other data-driven models, such as ANN and long short-term memory (LSTM), leverage big data and machine learning for flexible predictions. So, let us talk about the statistical snowmelt models. SWE-based regression is the most common statistical method that links snow water equivalent to total runoff using a linear regression model.



SW is measured manually along the survey line or automatically using the snow pillows with coefficients derived from observed data. Accurate sampling points are critical, and principal component regression helps optimize the location. This method can also predict spring peak flow and provide insight into soil water and groundwater conditions. The total snowmelt runoff (Q) can be estimated using a linear regression model, and here you can see the expression. So, here is what we are using: (a + b * SWE).

So, a and b are the coefficients derived from historical data through regression analysis. and SWE is the snow water equivalent that is measured at different depths. The characteristic curve is another method that uses a regression-based runoff characteristic curve relating spring flow cover ratios to runoff volume. These curves are unique to each basin but consistent across years, allowing forecasts using snow cover estimates. So, this is unique for the entire basin, and it is consistent with the air.

Snowmelt Modeling

Conceptual Snowmelt Models

- Conceptual snowmelt models estimate melt using a degree-day factor (DDF), which links temperature to melt rate through a linear relationship.

$$M = DDF \times (T_a - T_b)$$

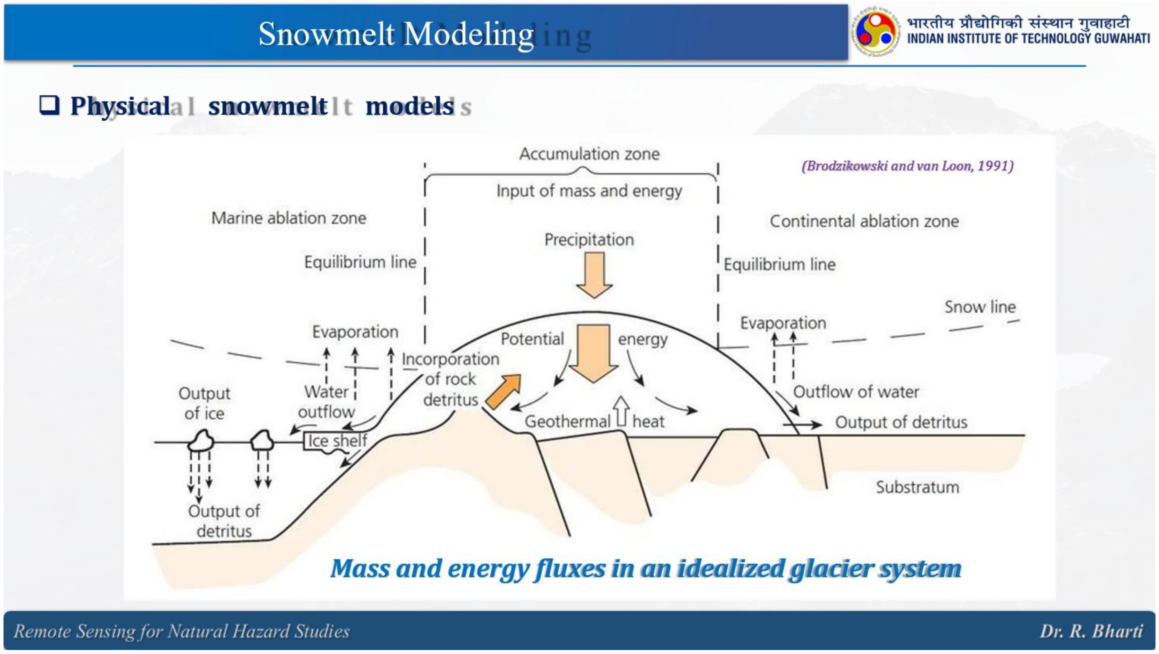
where T_a is the average daily temperature and T_b is the melt threshold temperature.

- **DDF** varies regionally and can be derived from measurements or empirical formulas.
- It can be adjusted using the ratio P_s/P_w, where P_s is snow density and P_w is water density.
- **Denser snow (higher P_s/P_w)** melts at a different rate than lighter snow, making local calibration important for accurate modeling.

Then comes the conceptual model. So, the conceptual snowmelt model estimates melt using a degree-day factor (DDF), which links temperature to melt rate through a linear relationship. So, here you can see the expression where, Ta is the average daily temperature and Tb is the melting threshold temperature. DDF varies regionally and can be derived from measurements or empirical formulas. It can be adjusted using the ratio Ps divided by Pw, where Ps is the snow density and Pw is the water density.

The denser snow where you have the higher Ps by Pw melts at a different rate than lighter snow, making local calibration important for accurate modeling. The limitations of the conceptual snowmelt model are that the traditional model assumes uniform ablation across the watershed, which is very rare often lumped or semi-distributed models. Spatial uncertainty arises due to the limited representation of local variations in snow and energy distribution. The constant daily degree-day factor does not account for intraday variations such as day-night differences.

Improved models include spatial and temporal variation, adding energy exchange terms for better accuracy. Then come the data-driven models. Data-driven models apply machine learning techniques to predict snowmelt runoff by identifying patterns in large data sets. Common methods include ANN, LSTM, SVR, GPR, and MLP, which are used globally in snow cover and runoff forecasting. These models can supplement physical models, improve parameter estimation, and help to fill data gaps like missing snow or rainfall records. Because, what happens when we talk about the glacier system is that some of the locations are impossible to visit, or it is very difficult to visit. So, sometimes even if you have satellite measurements, you need ground observations to calibrate them. So, if some datasets are missing, these models can help. However, they function as a black box, offering limited insight into physical processes and struggling with data scarcity and unseen conditions. Now comes the physical snowmelt model.

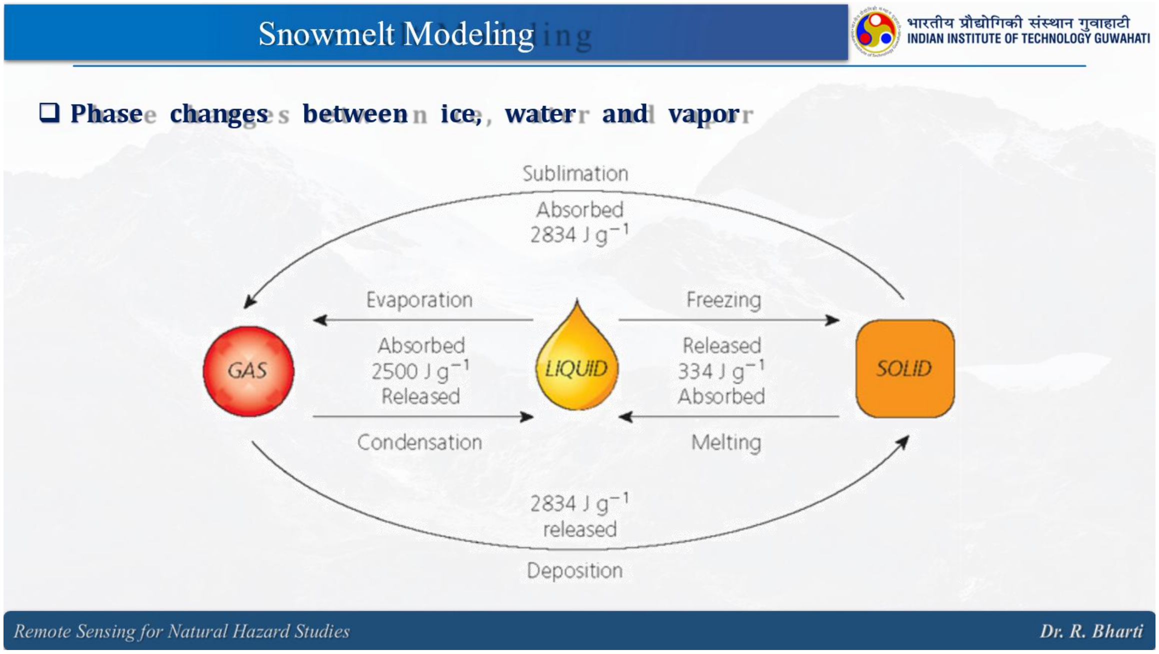


So, these estimates snow accumulation and melt using the energy and mass balance equation. So, now we will see the energy and mass balance equations. They include components like radiant flux, sensible and latent heat, and soil and rainfall heat fluxes to calculate melt rates. These models provide physically interpretable output, helping to assess the energy dynamics driving snowmelt in various conditions. So, personally, if you ask me, I will prefer to go with the physical snowmelt models.

They are suitable for complex terrain, though they require extensive input data, remote sensing aids in their calibration and spatial scaling. So, here you can see this is one of the examples of how the physical snow melt model works. So, here all the components will be incorporated into your model, and these will provide you with a very good insight into the glacier system. Their main challenge is scaling point-based observations to a large watershed due to data sparsity and parameter uncertainty. The empirical extrapolation methods are often used, though variability in terrain, vegetation, and microclimate remains a challenge.

With growing computing power, satellite data, and distributed models are becoming essential tools in snowmelt runoff forecasting. Now, let us see the surface energy balance. The energy exchange between the atmosphere and snow is ice. These can cause phase changes, specifically melting and freezing. This is what we are interested in because of the temperature change.

When snow ice is below 0 degrees celsius, added energy first raises its temperature; once at 0 degrees Celsius, any surplus leads to freezing. A net energy deficit can lower surface temperatures or cause ice deformation through condensation or freezing. Then comes the sublimation, which can also result in mass loss at subzero temperatures if atmospheric water pressure is low. Now let us understand this in detail. Melting and evaporation require energy to break hydrogen bonds, while freezing and condensation release energy.



This energy is called latent heat, technically referred to as the enthalpy of fusion or condensation. Melting requires 334 joules per gram, while evaporation or condensation involves approximately 2500 joules per gram. The total of all energy fluxes, the energy balance, determines changes in snow ice temperature and the glacier ablation rate. So, here you can see, and I hope this is very clear. So, suppose this is in the gas form. So, to convert gas to solid, you need to have 2834 joules per gram.

So, if it is in liquid, it is here to evaporate; you will require 2500 joules per gram and for freezing, you need 334 joules per gram that are being released. So, here this is evaporation, this is freezing from solid to liquid, and then from liquid to gas; then this whole thing, if you add this energy, will require 2834 joules per gram. I hope this is clear. So, this is called deposition; this is called sublimation.

Snowmelt Modeling

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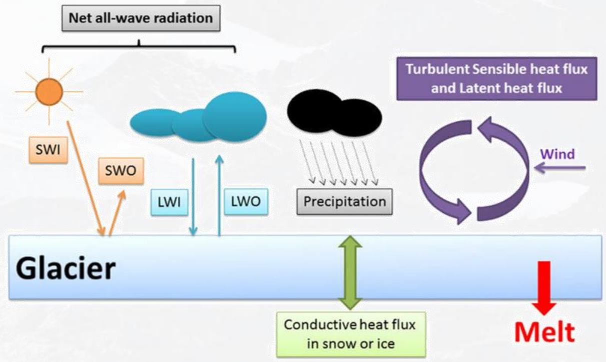
❑ Surface Energy Balance:

$$SW + LW + Q_H + Q_E + Q_R - Q_T - M = 0$$

where SW is net shortwave radiation flux, LW is net longwave radiation flux, Q_H is sensible heat transfer, Q_E is latent heat transfer, Q_R is energy from rain, M is energy used to melt ice or freeze water, and Q_T is energy used for temperature change in the ice.

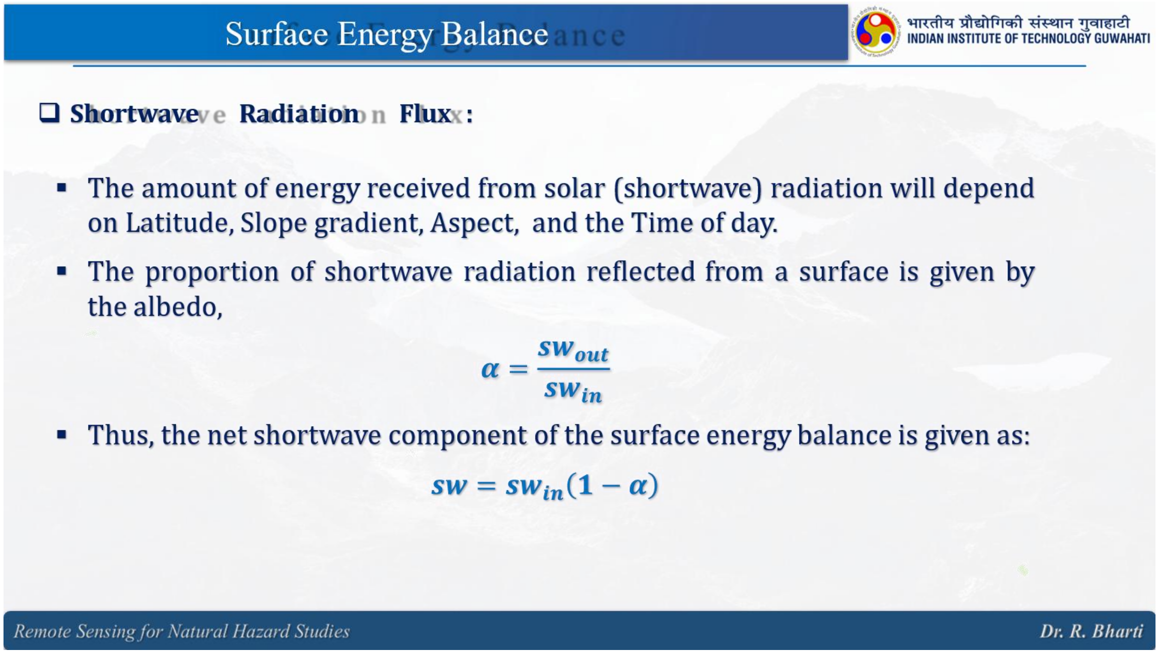


https://www.researchgate.net/profile/Mohd-Farooq-Azam/publication/285288613/figure/fig2/AS:614296119689235@1523471016480/The-processes-determining-the-energy-flux-at-the-glacier-atmosphere-interface.png

Remote Sensing for Natural Hazard Studies *Dr. R. Bharti*

So, the most important components of the energy balance at glacier surfaces are solar radiation, terrestrial radiation, and atmospheric longwave radiation. Sensible heat exchange with the atmosphere, latent heat transfer during condensation, evaporation, and sublimation, and heat supplied by rain, because the rain is also coming and bringing the temperature used to change the temperature of the ice. So, if you see the equation for the surface energy balance, So, here we are using the shortwave radiation flux, longwave radiation flux, sensible heat transfer, latent heat transfer, and the energy from the rain, which is brought to the system, and then the energy used to melt the ice or freeze the water; QT is the energy used for temperature change in the ice. So, here you see you have shortwave radiation, longwave radiation; this is incoming, outgoing, incoming, outgoing, and then precipitation will also bring the temperature into the system. Now, let us try to understand shortwave radiation flux and how it works.

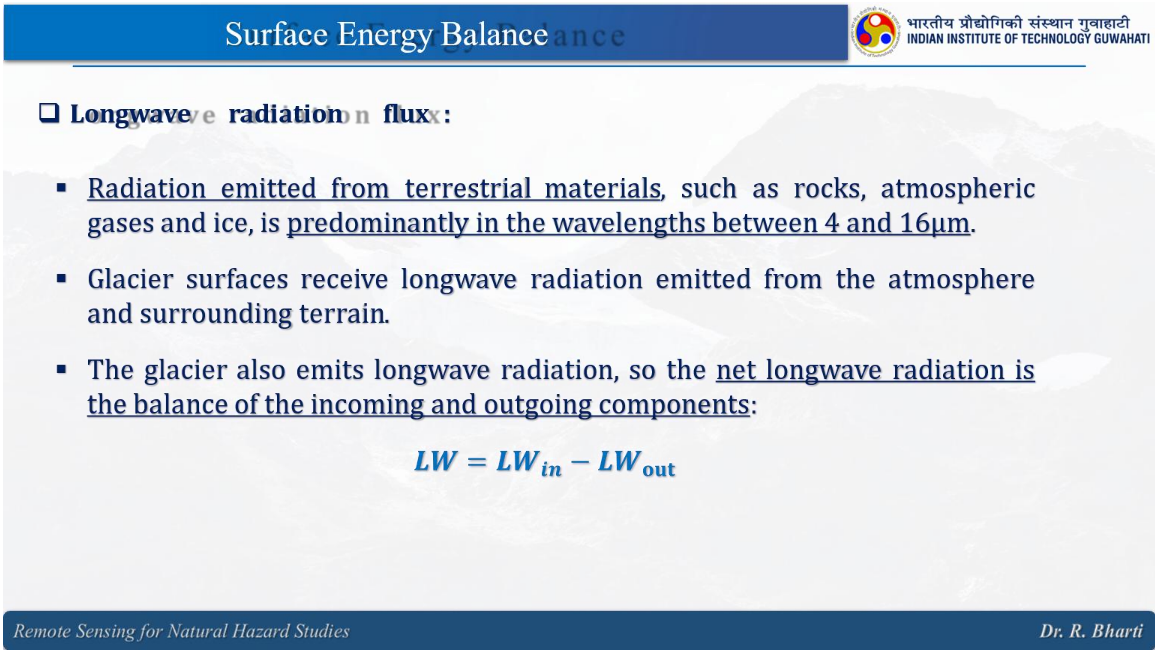
The amount of energy received from solar radiation will depend on latitude, slope gradient, aspect, and time of day. So, it is very simple to understand that the proportion of shortwave radiation reflected from a surface is given by the albedo. Now, let us see if this is the ice sheet; solar radiation is falling here. Then, if it is highly reflective in nature, that means the albedo is very high; a larger amount of energy will be equal to the incoming radiation and will be sent back to space. But if there is a mixture of some foreign particles here, if it is contaminated, what will happen? Albedo will be lower. In such condition, this will be greater than this one. So, here you will have more temperature in this system.



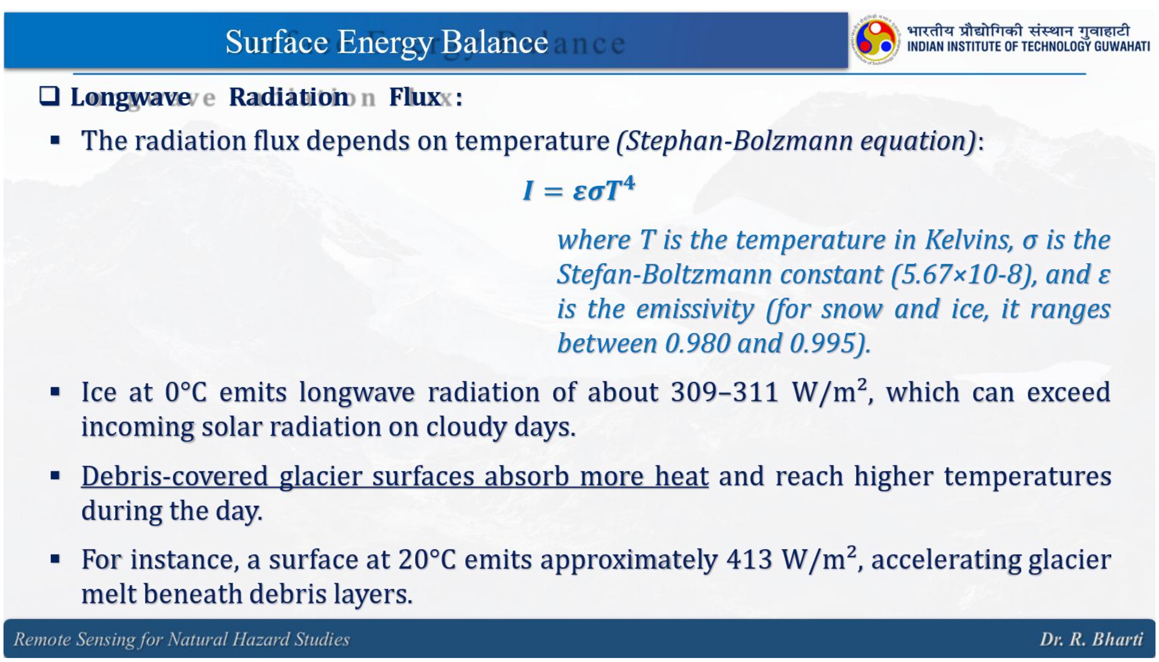
So here, the proportion of the shortwave radiation reflected from a surface is given by the albedo, which can be calculated. Thus, the net shortwave component of the surface energy balance is given by this expression. Then, fresh snow has a high albedo, as I said, and reflects most incoming solar radiation. Bare or dirty snow has a lower albedo, absorbing more energy and warming more quickly.

High albedo reduces the energy available for melting, preserving snow and ice surfaces. So, here you can see that this is the albedo a non-contaminated snow, and here it is contaminated. So, this is also a glacier system, but here it is fresh, and there it is contaminated. So, here you will have high albedo, here you will have low albedo.

If it has a low albedo, more temperature will be absorbed, and then you will have more melting. As snow cover diminishes during the melt season, the lower albedo on exposed ice increases the ablation rate, especially on the lower glacier slope. Then comes the long wave radiation flux; radiation emitted from terrestrial materials such as rocks, atmospheric gases, and ice is predominantly in the wavelength between 4 and 16 micrometers. The glacier surface receives long wave radiation emitted from the atmosphere and surrounding terrain.



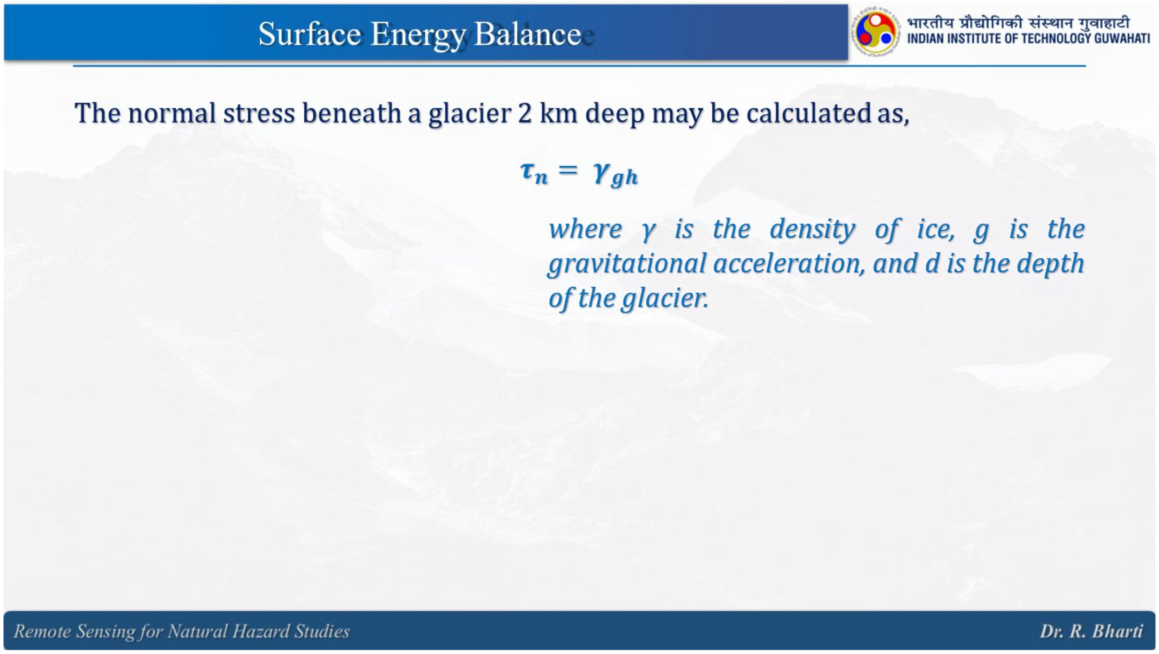
The glacier also emits long wave radiation. So, the net longwave radiation is the balance of the incoming and outgoing components. The radiation flux depends on the temperature. So, we will use or refer to this Stephen Boltzmann equation. where T is the temperature in Kelvin and sigma is the Stefan-Boltzmann constant, and here the emissivity is also used for the snow and ice, which ranges between this value and that value. Ice at 0 degrees Celsius emits long-wave radiation of about 309 to 311 watts per meter squared.



Which can exceed incoming solar radiation on cloudy days. Debris covering glacier surfaces absorbs more heat and reaches higher temperatures during the day. So, again, we are talking about some foreign material. So, the glacier system also has debris from the surrounding area because of various processes. Once you have the mineral and rock inside it, its thermal properties are different, and its behavior with respect to incoming radiation is different. So, they will absorb more light, or they can also reflect more, but most of the minerals and rocks absorb the energy, which will bring more temperature into the glacier system, and that will result in more melting. For instance, a surface at 20 degrees Celsius emits approximately 413 watts per meter squared, accelerating glacier melt beneath the debris layer. Now, we will have the sensible and latent heat flux. Sensible heat is the thermal energy transferred between the atmosphere and the glacier surface that causes a temperature change. It flows according to the temperature gradient from warmer air to colder ice or vice versa through molecular momentum. Latent heat involves energy exchange during phase changes such as melting, freezing, evaporation, or condensation.

Unlike sensible heat, latent heat does not change temperature but is crucial in deriving snowmelt, sublimation, and condensation on glacier surfaces. Then we will talk about the rain and heat transfer, specifically how the rain is affecting the temperature. When warm rainwater falls on a glacier, it transfers heat as it cools on contact with the cold snow or ice on the surface. If the ice is at 0 degrees Celsius, this energy contributes directly to melting. If below 0 degrees Celsius, then rain freezes, releasing latent heat that warms the ice. This process plays a significant role in modifying the glacier's surface energy balance, especially during rainfall events. The contribution of rain energy becomes more important in temperate glaciers and during warm-season precipitation. Then what are the different sources of heat in the glacier system? So, heat within a glacier originates from three main sources. The first one is solar radiation, then we have terrestrial radiation, and then we have basal sliding.

All three of these are the major sources. These energy inputs influence the thermal structure and internal processes of the glacier. Glaciers contain both temperate ice, which is at the pressure melting point, and polar ice. Cold ice remains below the threshold. The pressure melting point decreases with depth by roughly 1 degree Celsius for every 14 megapascal increase in pressure.



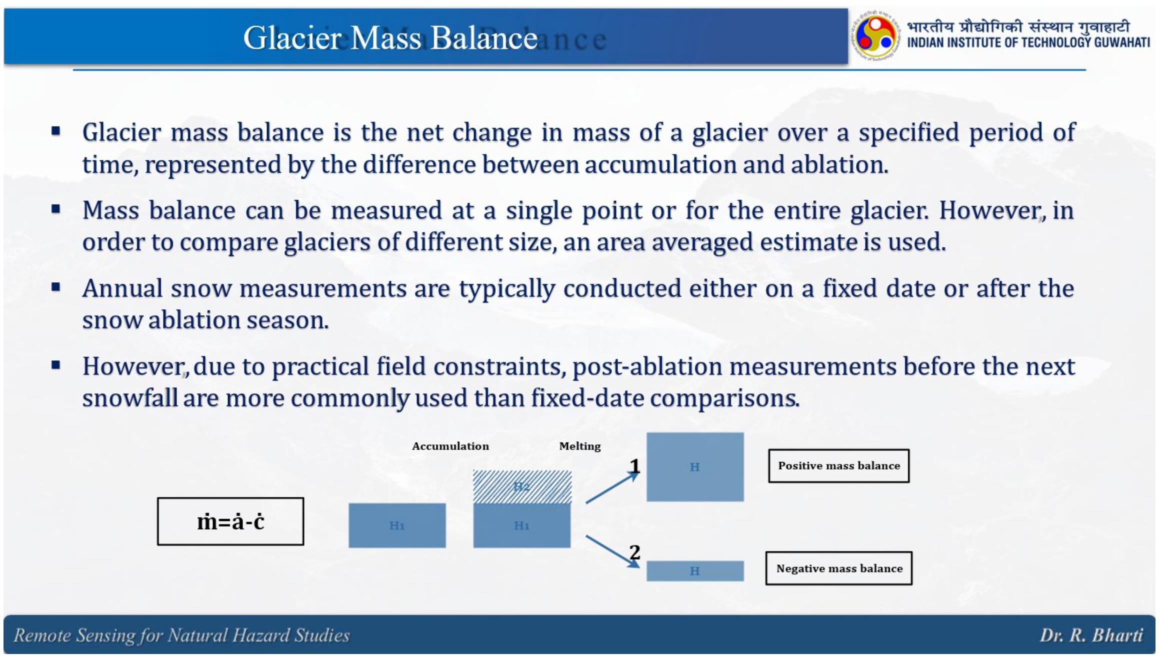
So, remember this relation. So, the normal stress beneath a glacier located at 2 kilometers deep may be calculated using this expression, where we are using the density of ice, then gravitational acceleration, and 'd' is the depth of the glacier for which we are trying to calculate the normal stress. Now, we will examine the basal heat transfer. So, as pressure increases with depth, you understood in the previous slide, right? So, as pressure increases with depth, there is a gradual temperature decrease from the glacier surface downward as we go down. Geothermal heat arises from the Earth's interior and reaches the base of the glacier through the bedrock. So, let us say this is the position where this glacier and the bedrock are interacting, right? Since heat flows from warm to cold, it cannot travel upward through the glacier's positive temperature gradient from cold to warm. Instead, this geothermal energy is used to melt the basal ice here.

So, geothermal energy is responsible for melting the ice at the contact position, converting it into latent heat and contributing to basal motion. So, now if you remember that now you put them this into the slope. So, because of this melting, the basal sliding will now start. In cold-based glaciers, temperature decreases upward from the base to the surface. This allows terrestrial geothermal heat to flow upward along the natural gradient from warm to cold.

The heat is efficiently conducted through the glacier body and released into the atmosphere because there is a gradient. So, it will from high to low. As a result, the interior of the glacier remains below the melting point, and the ice stays frozen to its bed. The glacier mass balance is the net change in the mass of a glacier over a specified period of time. It can be seasonal, it can be annual, it can be decadal, represented by the difference between accumulation and ablation.

Period or the amount represented by the difference between accumulation and ablation. Mass balance can be measured at a single point or for the entire glacier. However, in order to compare glaciers of different sizes and areas, an average estimate is used. So, here you can see what I have shown you in the previous lectures as well. So, if you have this kind of situation where you have an increase in the masses, we call it positive mass balance; if this is the situation, we call it negative mass balance.

Annual snow measurements are typically conducted either on a fixed date or after a snow ablation season. However, due to practical field constraints, post-ablation measurements before the next snowfall are more commonly used than fixed-date comparisons. Because, as I said, these are difficult terrains, it is very challenging to visit or revisit the area on a specific date. It depends on the climatic conditions, the approach, and the accessibility. So, not fixing the date if you can visit the area post-ablation is more ideal.



Snow accumulation and ablation occur at every point, with accumulation dominating at higher altitudes and ablation preventing at lower altitudes. At the equilibrium line, both processes are balanced. So, here you can see this is the equation, and here are the glacial subprocess and the process. Now comes the direct glaciological methods. So, these are very, very accurate. The direct glaciological methods involve direct field-based measurements of accumulation and ablation across the glacier’s surface.

Now, you see if this is the glacier's boundary. So, everywhere you have a stake that has the right measurement. So, you are taking the point measurement across this glacier. Accumulation is measured using snow pits or stakes marked annually, often identified by a dye layer. Ablation is measured by placing stakes into the glacier and recording how much ice melts around them.

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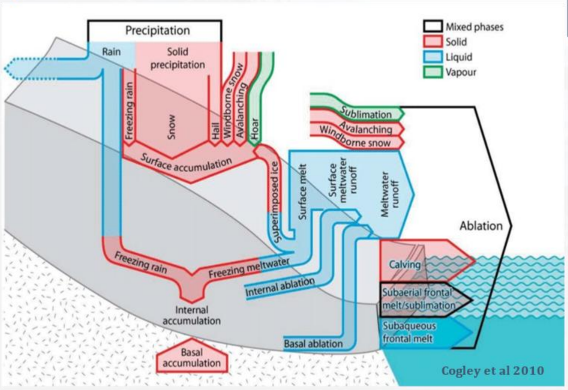


$$\dot{m}=(\dot{a}_{surface}+\dot{a}_{internal}+\dot{a}_{basal})-(\dot{c}_{surface}+\dot{c}_{internal}+\dot{c}_{basal})$$

Glacial Subprocess	Process
$\dot{a}$surface	Precipitation, avalanches
$\dot{a}$internal	Percolation + melt refreezing
$\dot{a}$basal	Freezing of subglacial water
$\dot{c}$surface	Surface energy balance comp
$\dot{c}$internal	Calving, frictional heat due to deformation, subaqueous frontal melt
$\dot{c}$basal	Basal abrasion and melting due to basal motion and geothermal heat

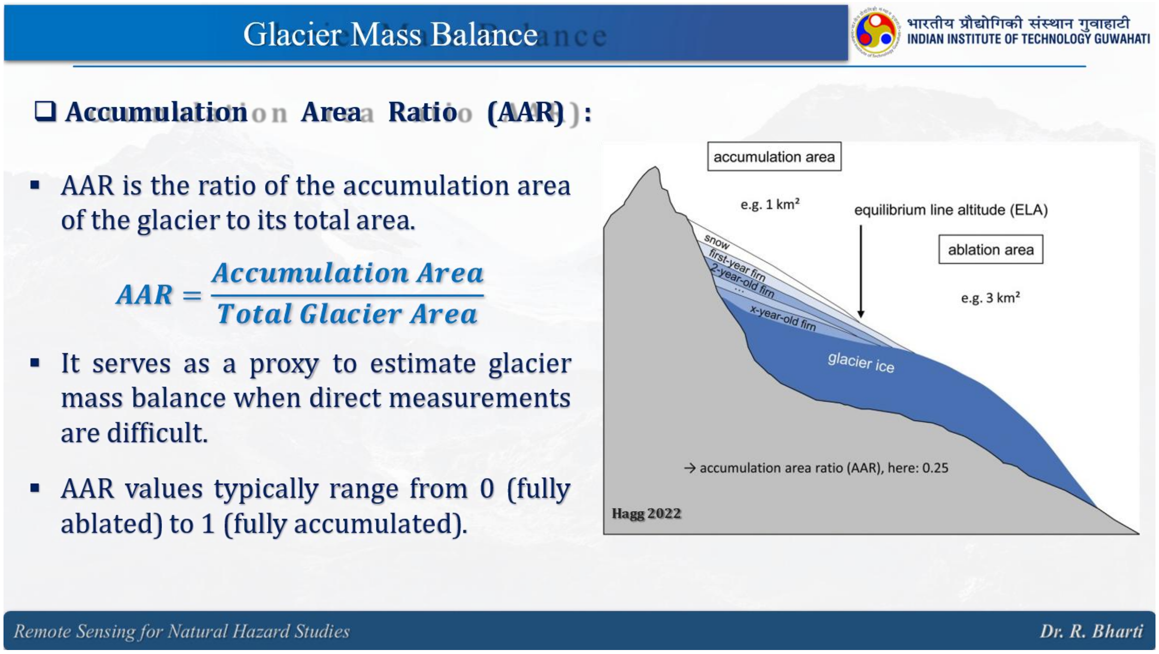
Point measurements are extrapolated. So, all these measurements will then be extrapolated. So, here it will be interpolated, but for this region, it will be extrapolated. To estimate glacier void mass balance. Density measurements are taken at each point to convert snow depth into water equivalents. Water equivalent is estimated using the thickness h, the density of water, and the average density of the annual layer.

So, that can be used to calculate the water equivalent. Now, if the sampling points are few or widely spaced, this procedure introduces a large error in the values. Though for this particular location, this measurement is very, very accurate, no other method can give you this information. But when we are doing this, this whole glacier mass is not approachable. So, we are doing some interpolation as well as the extrapolation. So, this will introduce large error into this result.

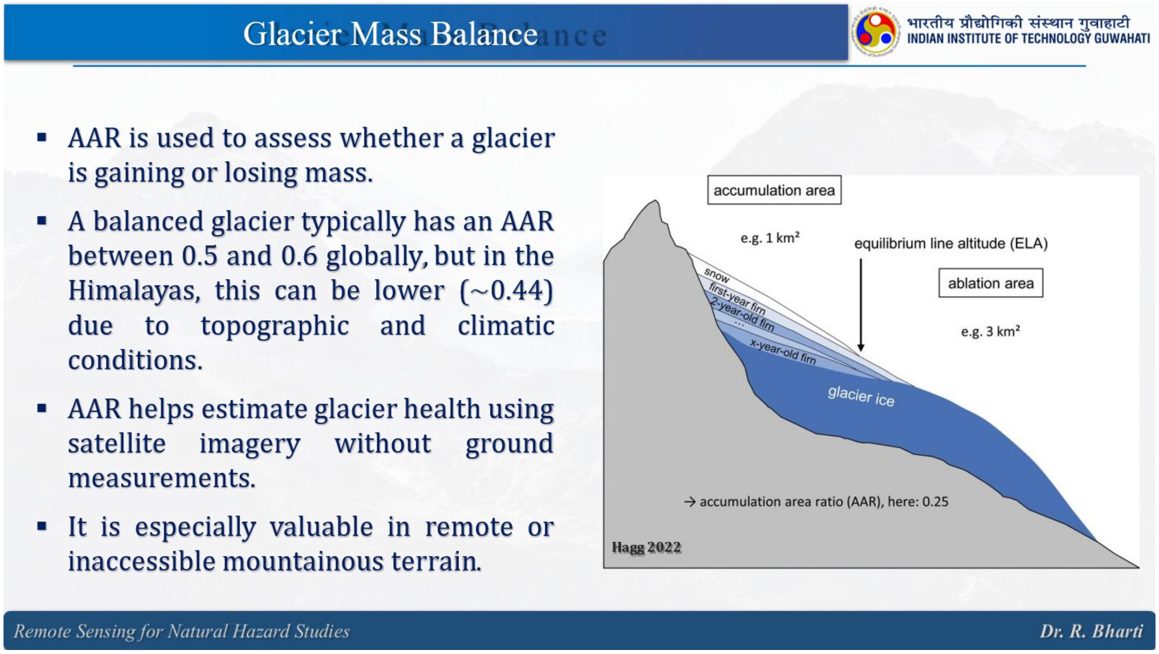
So, here you can see this is one of the examples. So, this is for the ablation stakes; then here you can see. So, here you can see in this figure that I have represented different points which have been used to measure the in-situ information for ablation and accumulation. So, this is from a paper. Now come the geodetic methods. The geodetic methods estimate glacier mass change by comparing elevation data from two different time points.

So, let us say this is a particular mountain where we have the glacier system. Now, what is happening when we have solid precipitation? It slowly gets accumulated, and then under pressure, this will be formed as glacier ice, and then again there will be accumulation on top of it, and then slowly it will be growing in size, and then slowly it will also move downward. So, if we have a digital elevation model for these two time frames, let us say this is 1980 and this is 2025. So, if we simply subtract it, we will be able to estimate what the difference in height is. Whether it is positive or negative can be easily calculated. It uses tools like DGPS, Lidar, and photogrammetry to detect surface elevation differences.

It reduces the need for dense in-field data collection by relying on satellite or aerial remote sensing data sets. Studies using this method help to understand long-term glacier volume change and spatial variability. So, here you can see there are two-time frames. We have the digital elevation model, and then this is the difference.



So that can be easily identified. Now, we have the accumulation area ratio (AAR), which is very popular. AAR is the ratio of the accumulation area of the glacier to its total area. So, AAR is the accumulation area divided by the total. It serves as a proxy to estimate glacier mass balance when direct measurements are difficult. AR values typically range between 0 and 1, where 0 is fully ablated and 1 is fully accumulated.



It is used to assess whether a glacier is gaining or losing mass, whether it is in a positive mass balance or a negative mass balance. A balanced glacier typically has an AAR value between 0.5 and 0.6 globally, but in the Himalayas, this can be lower, at 0.44, due to topographic and climatic conditions that we have observed in our work.

AAR helps estimate glacier health using satellite imagery without ground measurements. It is especially valuable in remote or inaccessible mountainous terrain. As I mentioned, these areas are very difficult to investigate. So, we have to look for proxy methods.

Now, we have the equilibrium line altitude (ELA). The ELA is the elevation on a glacier where accumulation equals ablation over the course of a year. It appears as a snow line at the end of the ablation season, which is in late summer. A rise in ELA suggests glacier retreat; a lowering ELA implies glacier advance. Like AAR, ELA is a key indicator of glacier mass balance. So, I hope you will be able to understand that this is the figure which beautifully explains the ELA.

Now, we will see the relationship between AAR, ELA, and mass balance. Now, we will try to link them to each other. AAR and ELA are inversely related; as ELA rises, that means less accumulation, AAR decreases, indicating mass loss. So, remember this: statistical models relate AAR and ELA to glacier-specific mass balance values. Both indicators can be derived from satellite images to infer trends in glacier mass change.

These are some of the advantages and limitations of the AAR and ELA methods. So, you can read this. With this, I will end this lecture.

Thank you. Thank you very much.