8/2/21

EC 2223 - LANDSCAPES

To study a peice of surface (topography)
- spatial scale
- how did it form and how is it evolving?
- What to measure and how?
  Map the surface / subsurface, forces, fluxes (material, energy, momentum)
- How to make models that'll provide quantitative explanations and predictions?

Concepts

→ Vertical movement - due to ~~mass~~ force imbalance : gravity, bouyancy, tectonic, elastic ....

Uplift - surface is rising, increasing altitude

Exhumation - altitude is steady, but lower levels of rock are coming to the surface

Subsidence - surface is sinking ⇒ decreasing altitude

→ Weathering and erosion (chemical & physical)

→ Mass transport / Deposition
- Viscous flow of fluids - mantle, ice, water, air
- Fracture of solids & gravity driven flow - landslides / earthquake

→ Formation / evolution is driven by processes that span across different components of the earth system, so an integrated approach is necessary

→
- Mass conservation
- Force balance
- Steady states
- Feedbacks.

Lecture 2 - Google Earth

* The Western Ghats have been formed due to continental rifting - when India broke away from the African plate. (Tarbuck Pg. 205, 206).
  Since then, the western Ghats have moved inward by a few kilometres and there are remnant mounts near the coast that haven't been eroded.

* Himalayan mountain ranges were formed when Indian and Euarasian plates collided. Even at the latitude of 27°N to 31°N, they have glacier caps at their highly elevated peaks.

In WG, rivers mainly shaped the landscape by eroding the rocks & ranges.
In the mountains, the glaciers shift and move to carry the sediments till the place ~~were~~ where they meet a river.

* Himalayas have been shaped by the flow of rivers - i.e fluvial landscape. If the sediment: water of river increases, in a plains, rivers can extend laterally. But in a mountain range it can only extend vertically.

* Sand dunes in the desert are shaped by wind. Their shape & scale is determined by
  - Direction of wind : multidirectional → star shaped
    unidirectional → transverse or longitudinal
  - Wind speed : sediment ratio

* Morphologically, the Alps are very similar - the are shape, the fluvial landscape, and rugged land
  This is because of they have similar genesis - convergence of two continental plantes.
  They've grown due to buckling & sediment deposition

Andes - longest continental mountain range - caused by the subduction of oceanic plate under the continental plate
Formed by volcanic activity.

When an ocean plate subducts under another oceanic plate, an island chain formed due to the mantle thats displaced upward by the subducting oceanic plate

Eg: Island arc ~~Kea~~ near Antarctica – equally spaced. why?

Pacific ocean plate subducting near the western coast of Japan – the volcanic activity causes earthquakes.

Mars

Primary geomorphic agent : wind

So we can find some dunes here, similar to the deserts

We can also find many craters on Mars. But their origin is highly debated because Mars doesn't fall in the crater (impact) window. So some suggest that its due to volcanic activity.

There are also some linear features on the surface which ig is generally attributed to faulting.

Google earth is not very accurate in measuring spatial scales on Mars due to lack of reference.

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Discussion

- A GIS (Geographic Information System) uses satellite imagery and wrap it around the globe.
- Each pixel is assigned a latitude and longitude using GeoControl points (GCPs) using georeferencing. A few GCPs are set apart to make sure error is minimal
- The elevation is taken care of using a geo elevation model and they are orthorectified (ie image as seen perpendicularly.

So these processes do result is some kind of error ie about 5m - 10m in horizontal scale

- To calculate elevation in mars and moon, you consider an equipotential surface and take that as your reference level.

Formation of Himalayas - faults formed due to compressional force that creates shear i.e. it goes up because there's nowhere else to go.

* Erosion plays an important role in the Himalayas - the monsoon dumps water on Himalayan slopes. This is an example of tectonic - climate feedback, because the monsoons intensified due to Himalayas and Tibetan plateau.

Another important example - Coenozoic cooling due to chemical weathering of glacier rivers.

* The shape of Sunderban delta is not typically triangular because its shaped by both fluvial and tidal flow.

3 types of deltas - Fluvial, Tidal and combination of the

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Lecture 3

Review of Classical mechanics

→ Newton's Law for a particle

In an inertial frame, the trajectory of a particle -

$$x(t) = x(0) + \dot{x}(0)t + \frac{1}{2}\ddot{x}(0)t^2 + \frac{1}{6}\dddot{x}(0)t^3 + \cdots$$

It can be written as power series of derivatives of $x(t)$ at $t=0$.

Newton simplified it to knowing just $x(0)$, $\dot{x}(0)$, $\ddot{x}(t)$

$$\Rightarrow \quad m\ddot{x} = \sum_i f_i(x, \dot{x}, t)$$

i.e. acceleration can be found out in terms of $x, \dot{x}, t$.

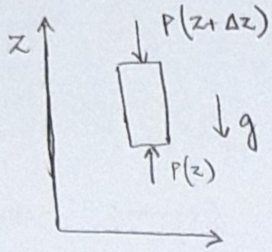
If there are driving and dissipating forces, at some point, they balance each other out and the particle reaches a steady state.

Particle at rest : $\sum_i f_i = 0$

If particle is in steady equilibrium, it will go back to that state if there are small perturbations.

Equilibrium need not be steady

Fluid at rest



For a fluid at rest,

$$P(z).da - P(z+\Delta z).da^{\#} - \rho g.da.dz = 0 \quad -(1)$$

$$\Rightarrow \partial_z P + \rho g = 0$$

\# Taking Taylor series expansion and ignoring the higher order terms

Taylor Series : $f(x+h) = f(x) + h f'(x) + \frac{h^2}{2} f''(x) \cdots + \frac{h^n}{n!} f^n(x)$

$$P(z+\Delta z).da = P(z)\,da + dz\,P'(z)$$

Eq^n (1) becomes —

$$\cancel{P(z)}da - \cancel{P(z)}da - P'(z).dz\,da - \rho g\,dz\,da = 0$$

$$\therefore P'(z) + \rho g = 0$$

$P'(z) = \partial_z P$: Vertical pressure gradient

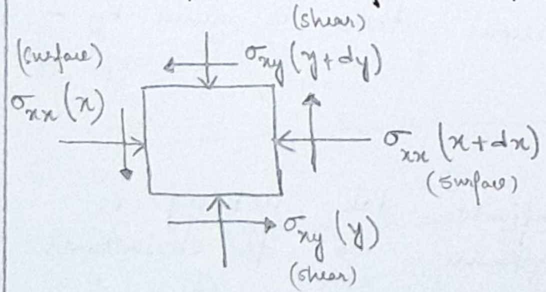
$$\frac{dP}{dz} = -\rho g \Rightarrow \int_{P(0)}^{P(0+z)} dP = -\rho g \int_0^z dz$$

$\therefore P(z) = \rho g z$ if fluid surface is at $z=0$ & $P(0)=0$

(?) Example of fluid in an unsteady equilibrium?

Solid rest

We can apply force perpendicular and parallel to the surface of the solid.



The translation (σ_{xx}) and shear (σ_{xy}) forces both have to be balanced for the solid to be at rest.

σ ; stresses

$$0 = \sigma_{xy}(y)\,dx\,dz - \sigma_{xy}(y+dy)^{\#}.dx\,dz + \sigma_{xx}(x).dy\,dz - \sigma_{xx}(x+dx)^{\#}dy\,dz$$

$$- \partial_y \sigma_{xy}(y).dx\,dy\,dz - \partial_x \sigma_{xx}(x).dx\,dy\,dz = 0$$

$$\therefore \partial_y \sigma_{xy}(y) + \partial_x \sigma_{xx}(x) = 0 \quad : \text{X-axis}$$

$\hookrightarrow$ (1)

⑥

$$\sigma_{yy}(y)\cdot dx\cdot dz - \sigma_{yy}(y+dy)\cdot dx dz + \sigma_{xy}(x+dx)\, dy dz$$
$$- \sigma_{xy}(x)\cdot dy dz - \rho g\, dx dy dz = 0$$

$$\sigma_{yy}(y)\, dx dz - \sigma_{yy}(y) dx dz - \partial_y \sigma_{yy}(y)\cdot dx dy dz - \rho g\, dx dy dz$$
$$+ \sigma_{xy}(x)\, dy dz + \partial_x \sigma_{xy}(x)\, dx dy dz - \sigma_{xy}(x) dy dz = 0$$

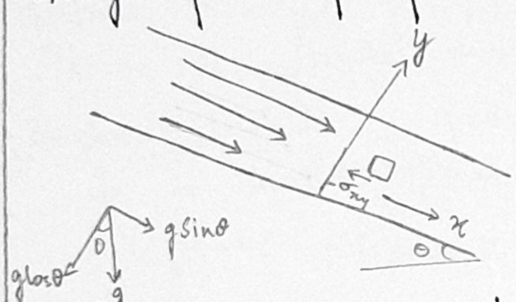
$$\partial_x \sigma_{xy}(x) - \partial_y \sigma_{yy}(y) - \rho g = 0 \quad \text{—} ②$$

Y-axis.

Argha's eq^n : $- \partial_x \sigma_{xy}(y) - \partial_y \sigma_{yy}(x) - \rho g = 0$

According to this, if there are no shear forces, then vertical pressure gradient is balanced by gravity : lithostatic pressure.

Steady flow of fluid on a infinite slab



Even though the liquid is flowing down at an incline, its not accelerating because gravity is balanced by viscous force. Viscous force is created because of friction between two layers which are moving at different drift velocity

So velocity of different layers of the fluid changes with y i.e. $v_x(y)$

Shearing b/w layers i.e. viscous stress is given by -

$$\sigma_{xy} = -\mu\, \partial_y v_x$$

$\mu_{water} = 10^{-3}$ Ps

$\mu_{air} = 10^{-5}$ Ps

As we're considering an infinite slab, nothing is changing along x-axis. So ~~its~~ derivatives wrt. x-axis are 0.

The equations we get are similar to ① and ② except ∂_x terms are 0 and the rotation comes in as sinθ / cosθ.

$\therefore$ We have —

$$-\partial_y \sigma_{xy} + \rho(+g\sin\theta) = 0$$

$$-\partial_y \sigma_{yy} - \rho g\cos\theta = 0 \quad \text{— Hydrostatic (vertical)}$$

Enough to consider in this eq^n?

From previous equation, $(-)\mu\, \partial_y^2 V_x + g\sin\theta = 0$

We get this by taking boundary conditions that $V_x(0) = 0$ and $\partial_y V_x = 0$ at surface

Do this integral!

From this, we get —

$$V_x(y) = \frac{g}{\mu}\sin\theta\left(\frac{y^2}{2} - Hy\right)$$

$$\frac{d^2 V_x}{dy^2} = \frac{g}{\mu}\sin\theta \Rightarrow \iint d^2 V_x = \iint \frac{g}{\mu}\sin\theta\, dy^2$$

$$\int_{V_x(0)}^{V_x(y)}\int d^2 V_x = \int_0^y \int \frac{g}{\mu}\sin\theta\, dy^2 = \frac{g}{\mu}\sin\theta \int_0^y (y+H)\,dy$$

$$\int_{V_x(0)}^{V_x(y)} dV_x = V_x(y) - 0 = \frac{g}{\mu}\sin\theta\left(\frac{y^2}{2}\Big|_0^y + Hy\Big|_0^y\right)$$

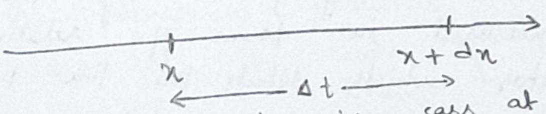
$$\therefore V_x(y) = \frac{g\sin\theta}{\mu}\left(\frac{y^2}{2} \oplus Hy\right) \; ?$$

(Mass) Conversation

Source strength: no. of cars added at some particular point x. — $s(x)\cdot dx$

$$Q_x(x+dx) = n(x+dx)\, V_x(x+dx)$$

$$Q_x = n(x)\, V_x(x)$$



n : no. of cars.

$Q_x(x)$: Flux of incoming cars at x

$Q_x(x+dx)$: Flux of outgoing cars at $x+dx$

If no. of outgoing is different than no. of incoming cars, then change in no. —

$$\Delta x\, \Delta n = [Q_x(x) - Q_x(x+dx)]\, dt + s(x)\cdot dx\, dt$$

Dividing by '$dx\, dt$' on both sides

$$\partial_t n = -\partial_x Q_x + s(x)$$

Lecture 04

Digital Elevation Model (DEM)

DTM - Digital Terrestrial model   DSM - D. Surface model

The information about the elevation of the earth's surface can be stored in a matrix that is DEM

Its the most common digital data of shape of earth's surface

Raster presentation - in grid format whose basic unit is grid

Vector formats - Contour line model : contour lines connect points of equal elevation

Triangular irregular network - useful for spatial analysis. Each node has certain elevation value Then, triangles are drawn after some geometric constraints

Any remote sensing model will give us DSM - ie the surface of the landscape

DEM Acquisition

- ★ Ground survey (Auto-label [more labour], total station etc)
- ★ Digitisation of topographic contour lines.
- ★ Differential GPS - very accurate, high resolution, vertical uncertainty is very less
- ★ Stereo photogrammetry - If we have optical images of overlapping areas, this method can be used to develop DEM
  ISRO product - cartosat
- ★ LiDAR - Light Detection and Ranging scanning. Points on surface are measured in terms of relative distance from the detector which detects the time taken for laser to reflect. Generates a very dense point.
- ★ Radar interferometry - gives DEM on a very large scale with moderate resolution. This is done using SRTM.
  (Shuttle Radar Topography Mission)

Usage of DEM –

- Slope of surface : controls the flux of different elements in the landform (gradient of a surface)
- Aspect : slope direction – downslope direction of maximum rate of change in value from each cell to its neighbors.
  - → Azimuth of any slope facing – controls geomorphological factors etc.
- Identification of geological structures
  With Radar interferometry – regional structures like faults, folds (~ 10s of km) can be identified
- 3D simulation
- Change analysis – calculating erosion / pollution of any region at high temporal and spatial resolution
- Derive contour maps
- Hypsometry – tells us the fraction of area below/above certain elevation – used for quick landscape characterisation (young/new ? Evolving?)

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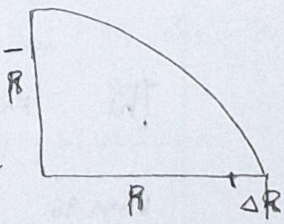
Lecture 06

Avg g and ρ are considered

Oblateness of a Rotating Planet

In the equatorial direction –

Pressure at the centre of earth due to –



- Centrifugal force : $-\int_0^{R+\Delta R} \rho \omega^2 r\, dr \approx -\int_0^{R+\Delta R} \rho \omega^2 r\, dr$

$$\approx -\rho \omega^2 \frac{R^2}{2} \quad \text{—(1)}$$

- Weight of bulge : Due to hydrostatic equilibrium, $P_p = P_e$

So, $\int \rho g\, dr = \rho g \Delta R$ —(2)

ρ increases with depth ⇒ earth is a stratified sphere

Equating ① and ② —

$$\rho g \Delta R = \frac{\rho \omega^2 R^2}{2} \Rightarrow \frac{\Delta R}{R^2} = \frac{\omega^2 R \rho_{avg}}{2 g \rho_{crust}} = f \text{ i.e flattening}$$

$$f = \frac{\Delta R}{R} = \frac{(2\pi/86400)^2 \cdot 6.4 \times 10^6 \times 5.5 \times 10^3}{2 \times 9.8 \times 3 \times 10^3} \approx \frac{1}{298} = 0.003$$

$\therefore \Delta R \sim 22$ km   # largest topographic feature on earth.

(Order of magnitude calculation).

The formula of f is generally applicable to all rotating planets. Here, we've balanced hydrostatic equilibrium, not the forces. (assuming earth to be a spherical liquid in hydrostatic balance).

Oblateness grows with ω^2 and difference b/w ρ_{avg} & ρ_c

We can see that oblateness of atmosphere is much greater.

Some planets (Mercury, moon) have different flattening than estimated. This could be because of greater tidal forces they face

Topography

The earth is not a uniform stratified spheroid. Because of differences in temp, density and mantle activity and mainly uneven distribution of mass

There are a lot of deformations are irregularities on the geoid (shape that ocean surface would take under the influence of gravity and rotation alone, in absence of winds and tides).

The maximum depression due to lack of density mass is about 100 m. This occurs in the Indian ocean.

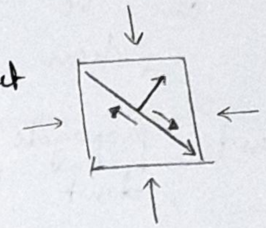
EGM 96 - commonly used

Undulation - height of the geoid relative to a given ellipsoid of reference

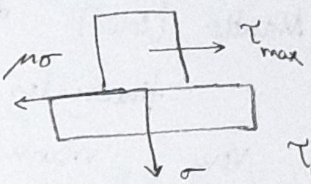
Maximum elevation

Consider a mountain range on planet (M, R, ρ). To estimate the maximum elevation possible, we need to know the stress limit of the material

There are no shear forces acting on a body in lithostatic equilibrium. But when load is increased on one side, shear forces / stresses develops in the interior of the object along any plane and at some point, it'll fail.

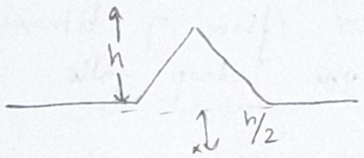


Frictional strength increases linearly with pressure, upto a limit



At ~ 1 GPa, plastic failure occurs

$\tau_{max} \sim 0.85\sigma$



Maximum stress difference is experienced at a distance h/2 below surface

In reality crustal failure happens at 100 MPa

$$100 \text{ MPa} \approx \rho_c g \Delta h_{max}/2$$

$$\therefore \Delta h_{max} = \frac{2 \times 10^8 \times R^2}{\rho_c GM} = \frac{3 \times 10^8}{\rho_c G (2\pi \rho_{avg})} \cdot \frac{1}{R}$$

So, $h_{max} \propto \frac{1}{R}$

There is a discrepancy with moon : Expected ~49, obs - 8.7

This is because there are no tectonic or volcanic activities to create such huge structures.

Lec 7

Hypsometry – distribution of elevations – shows 2 modes

~ 2 km on continent – low density

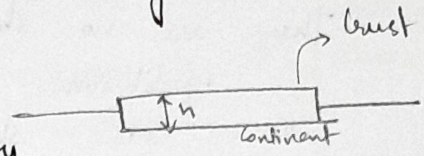
~ 4 km below in ocean basin ⇒ higher density

Assume everything's floating on mantle. Hydrostatic eq.

Mantle – not actually liquid.

can be proved using shear waves in seismology

can't propogate through liquids ∵ they don't have shear strength



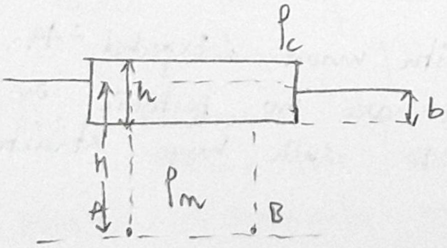
Mantle (solid) due to defects can flow over longer timescales v. slowly

Very viscous – $10^{25}\eta_{air} = 10^{22}\eta_{water}$

Pitch drop expt – very v. slow flow of bitumen. Actually a solid, but one drop falls every 8-9 years.

⇒ Solid materials under stress (mantle – thermal + gravity) can creep & flow v slowly.

Floating crust emerged from "missing mass" in geographical surveys due to low densities.



Crust floats in hydrostatic eq. over mantle. by buoyant forces.

b: compensation depth: going down vertically, where does density make a sudden transition

At any depth below b, pressure should be balanced and equal at 2 points at the same depth – $P(A) = P(B)$

Set $H = b$. Then,

$$g\rho_c h = g\rho_m b \quad \text{(pressure balance)}$$

Height of crust: $h - b = \frac{\rho_m b}{\rho_c} - b = h - \frac{\rho_c h}{\rho_m}$

$$\therefore \Delta h = h - b = h\left(1 - \frac{\rho_c}{\rho_m}\right)$$

Principle of isostasy:

State of gravitational eq. between crust & mantle is st. crust floats on the mantle at an elevation dependent on its thickness (h) and density (ρ_c).

Same eq^n works for any H -

$$\rho_c h + \rho_m (H - b) = \rho_m H \Rightarrow \rho_c h = \rho_m b$$

Less density (ρ_c)
More thickness (h) } more crustal elevation

ΔH - how much of the crust sticks out

$\rho_m = 3.3\ g/cm^3$ (using volcanism)

$\rho_c = 2.7\ g/cm^3$

$$\Delta H = h\left(1 - \frac{2.7}{3.3}\right) \simeq \frac{2h}{11}$$

h is typically of the order of 40 km (continental crust) $\Rightarrow \Delta H \simeq 6$ km

Oceanic crust has different ρ and h. h is very low $\Rightarrow \Delta H = 0$

On average, continental crusts are 6 km higher than oceanic depths

Consistent with bimodality of histogram

Mean sea depth - 4 km

2 km continental elevation is approx — closer to 0.9 km in reality — still v. close estimate for such a crude model — same order of mag.
∃ density variation with depth & h varies across the slab

By isostasy, mean continental elevation is 2 km
Close correlation b/w crustal thickness & elevation distribution.

Mountainous areas — subduction, one plate going under another. h is very high ∵ 2 crustal layers
$\Delta H \propto h$
High elevation due to isostatic adjustment

Continental elevation — sharp peaks
Oceanic depth distribution very wide
Possible sources: differences in densities between oceanic and continental material.

oceanic ↓ Fe/Mg rich basalt, mafic ~ 3 g/cc
h is low

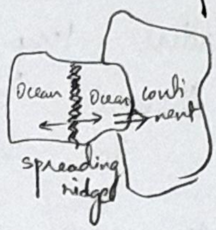
continental ↓ granitic - more Na/K/Al - felsic ~ 2.7 g/cc
h is more.

Broad peak of the curve ⇒ crust thickness or density (or both) of oceanic crust may vary over large ranges.

Ocean basin —
∃ oceanic ridges - hotspots of earthquake activity
Eg. mid-atlantic ridge
long mountain ridges — volcanic clusters, continuous, new ocean floor is made here from sea floor spreading (tectonic activity)

Divergent plate boundary → Seafloor spreading
mantle upwells to form new seafloor

Plate tectonics happens because of subduction of associated oceanic plates to continental plates

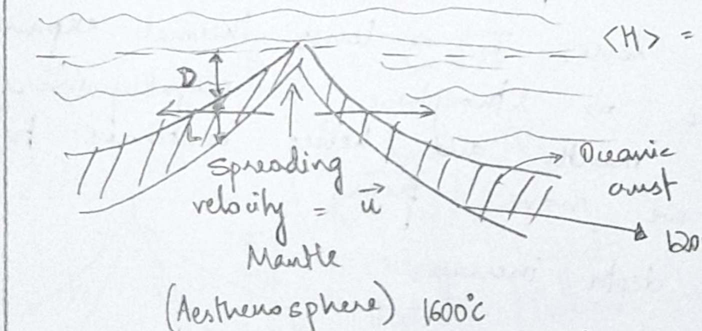


Subduction needn't happen — but spreading always does.

Total amount of ocean floor stays the same because Σ spreading & Σ subduction balance each other out

Newest basalt — oceanic, oldest — near continent
Nothing older than ~ 300 ma — period of Wilson cycle
By then it becomes so compacted → more dense than mantle that it subducts
gravitational instabilities knock it out.

PLATE COOLING MODEL



⟨H⟩ = 2.5 km below sea level

Thin Oceanic crust forms from mantle cooling

Younger crust : highest elevation

Due to this basalt furthest away from mid ocean ridge (MOR) will be most dense and thickest

Mean T (lithosphere) ~ 600°C

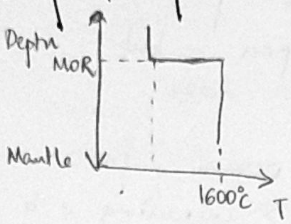
Lower boundary of lithosphere is isotherm of 1200°C
↳ goes deeper as we move further away from MOR

Lithosphere is homogenous – no melting, only thermal creep
So, at any point P, we know that

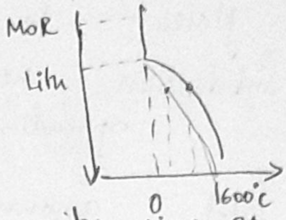
age of lithosphere = $\frac{\text{dist (MOR, P)}}{\text{velocity } u}$

Vertical temp. profile
Assume ocean water is 0°C (works as approx ∵ 600° is compairitively huge)

At MOR :

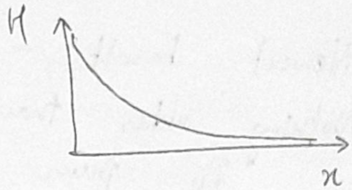


Away



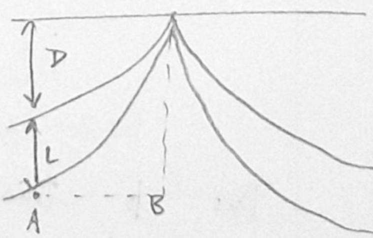
Either with time as seafloor spreads or actually move some distance away

With time, the 1200°C isotherm slides down with distance in a 'decaying' manner
T profile allows conduction as seafloor spreads



Suppose at T, the seafloor at MOR at t=0 has the thickness of lithosphere as L.

Because of ~~tt~~ lesser T ⇒ lesser thermal expansion,
both basaltic ⇒ lithosphere is slightly more dense than mantle and hence continues to sink as we move from MOR, so ocean depth increases



Lithosphere thickens & becomes denser with time

Amount of sinking regulated by isostasy –

$$\rho_m (D+L) = \rho_{water} D + \rho_{Lith} L$$

We can find thickness at given distance using heat conduction eq^n — assume 1D Heat loss vertically to ocean —

$K : m^2/s$

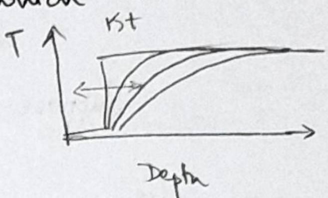
$$\partial_t T = \partial_x^2 (T) \cdot K \quad : \text{Diffusion eq}^n$$

↳ diffusion constant

Find L from RHS, total height from which heat is lost.

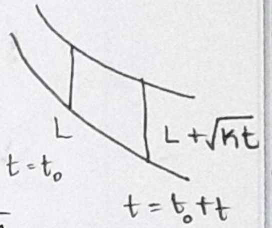
$K \cdot t$: (Diffusion length)^2

It controls the length over which a "significant" T. change has occured length.



$\sqrt{Kt}$: Diffusion

Its the avg 'progogation length' of a region after time t



$\sqrt{Kt}$ is the rate at which L increases with time

Assume order is one. Approx,

$L(t) = \sqrt{Kt}$ i.e lithospheric thickness $\propto \sqrt{age}$

Then, $\rho_m D + \rho_m \sqrt{Kt} = \rho_w D + \rho_L \sqrt{Kt}$

$$D = \left(\frac{\rho_L - \rho_m}{\rho_m - \rho_w}\right) \sqrt{Kt}$$

D_0 is the ridge height which maybe underwater at D_0 depth [$\rho_w D_0$ cancels on both sides]

$$\therefore D(t) = D_0 + \left(\frac{\rho_L - \rho_m}{\rho_m - \rho_w}\right) \sqrt{Kt}$$

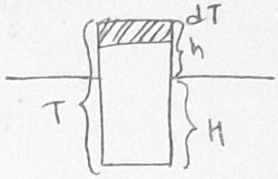
$D_0 \sim 2.6$ km $\quad \rho_L = 3.39 \quad \rho_m = 3.3 \quad \rho_w = 1$

$t = 100$ Mya (avg age of lithosphere)

$\Rightarrow D(t) = 4.7$ km $\Rightarrow$ Mean ocean depth ~ 4 km despite crude calculation

Exp data agrees pretty closely with theoretical ones
So, over a large range, $\sqrt{Kt}$ works
Conductive cooling thickening the ocean lithosphere and sinking probably plays a big role
Deviations – mantle convection changing for older ocean floor – cooling rate changes when older because of heating from mantle

Surface processes can also change crustal thickness
Eg: Non-uniform erosion.
Suppose we start with thickness T, $\Delta H = h$ and $T - h = H$
Suppose a block of height dT is removed from area above sea level



$(\text{Total height})_{new} = T - dT$

Sediments get eroded and transported to the oceans.
Sharp peak at 0 in hypsometry caused by sediments – & creates plain along the coast and deposit flatly
Carried sediments hit a boundary while in rivers
When they meet ocean, rivers have no KE, there exists no slope to the land, already at surface, while wholly enriched with sediments – cannot carry. Must deposit to form coastal plains.
Maintained by sea erosion
Erosion rate depends on topography: can be few mm/yr to m/yr
Assuming no new things are being added, T decreases –

$$\Delta H = T\left(1 - \frac{\rho_c}{\rho_m}\right)$$

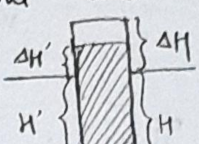
$$\Delta H' = (T - dT)\left(1 - \frac{\rho_c}{\rho_m}\right)$$

} Change in height – $dT\left(1 - \frac{\rho_c}{\rho_m}\right)$

$\Delta H' < \Delta H$ → eroded crust sticks out less

Also, sinks into the mantle less

∴ lesser weight pushing it down.



$H' < H$ change : $dT \left(\frac{\rho_c}{\rho_m} \right)$: crust bottom moves up by

Decrease in height above the mantle — $dT \left(1 - \frac{\rho_c}{\rho_m} \right)$

Elevation change above mean sea level

Note: If dT gets eroded away, it's not that ΔH decreases by dT; elevation of surface dips by an amount < dT because of isostatic compensation.

Bottom of crust moves up by k_1 and ~~bottom~~ top of crust moves down by k_2

$k_1 + k_2 = dT$

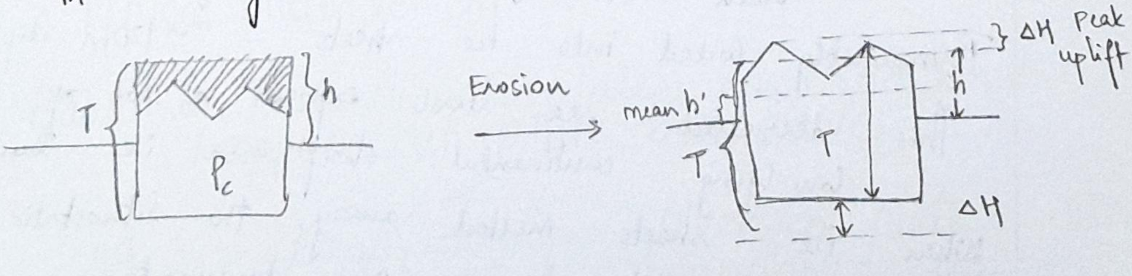
Hence, erosion can cause isostatic uplift of crust bottom (& not affect elevation much)

Erosion leads to more inner layers of crust and brought to the surface from deeper crustal layers.

But it doesn't remove layers linearly but in blocks.

Strength of erosion agents & erodability of material varies

While mean elevation declines due to loss of crustal mass in erosion, greater erosion in valleys can cause peaks to uplift.



Mean elevation change : $h - h'$

Mean elevation change is same for when dT is removed, but isostatic compensation at the bottom pushes the peaks up by $dT\left(\frac{\rho_c}{\rho_m}\right)$

This is ∵ T between base and peak is preserved as it remains uneroded: as IC pushes base up by ΔH, peak moves up by ΔH over the original position, though mean height falls by $dT\left(1-\frac{\rho_c}{\rho_m}\right)$

∴ Differential erosion can increase elevation.

Also creates sharper differences in relief by making deep valleys — these are where lithosphere load is removed to cause IC.

<Height> at surface dips by $\langle dT\rangle\left(1-\frac{\rho_c}{\rho_m}\right)$ but high elevations can be made — isostasy shapes topography

---

Another eg. of IC and topographic change: Large scale movement of water

Basically: change lithospheric load at surface somehow

Earth in Pliestocene (~20 kya): LGM - lots of large ice sheets - landlocked ice sheets NA, Asia, Europe

⇒ Ocean levels ↓ ∵ ↑ water sequestered

Permanantly locked into ice ~~sheets~~ — ~120m dip in sea level

This decreased sea level exposed a lot of low-lying continental shelf area i.e. land % ↑

When ice sheets melted away, the lithostatic load ↓ so the crust became more buoyant

⇒ IC : land surface goes up in elevation.

Lecture

Weathered and erosion
↳ Break-up of material ; Transport of broken up material

Denudation : elevation of landscape decreases
This is the result of erosion
It exposes sub-surface level rocks

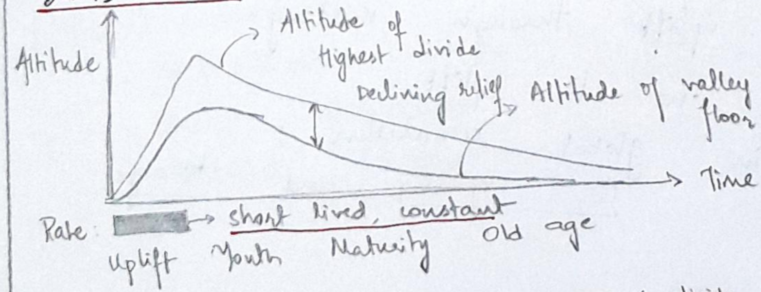
Landscape Evolution
Uplift – adding to the lithosphere
Erosion : Exhumation : exposing rocks
Denudation : removing debris & placing elsewhere

The newer school of thought subscribe to the idea that there two processes and closely linked and their interplay affects the evolution of landscape

Models of Landscape Evolution
Most models consider mountains to build & explain their

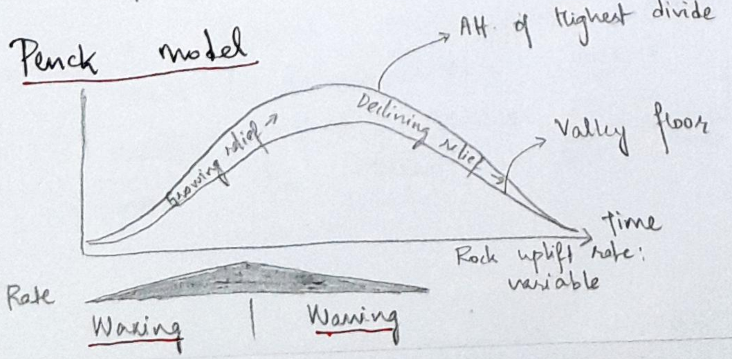
Relief : difference b/w highest & lowest point

1. Davis model

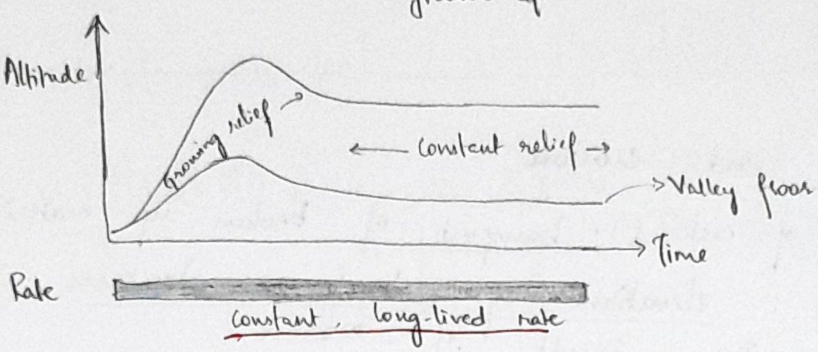


2. Penck model

Most realistic model
There's a cyclicity in uplift

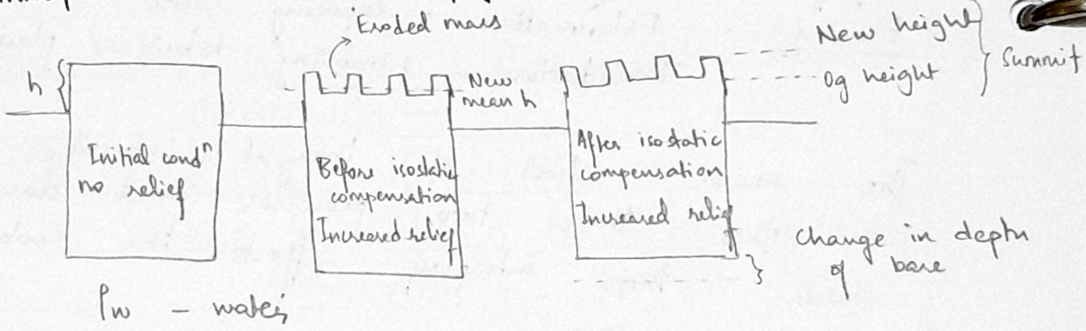


3. Hack model



Erosionally Driven Summit Uplift

This is based on isostatic balance and compensation and brings up the fact that summit can be uplifted because of erosion.



New insights

- High mountains affect weather pattern
  Increase in precipitation
  Glaciation
- Erosion drives uplift through isostacy
  due to uplift
- Climate change in global circulation;
  Change in CO_2 cycling and storage

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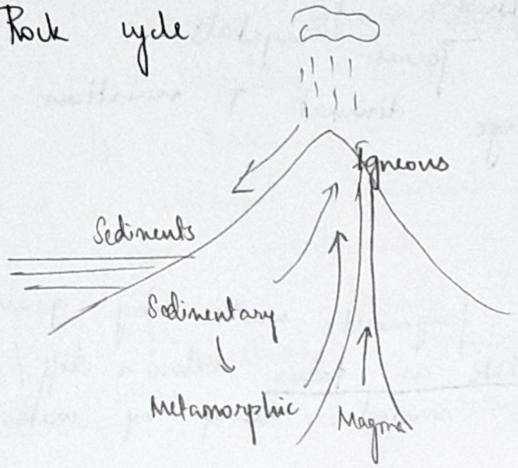
Lecture 12 - Weathering

It breaks down the rocks to smaller peices so they can get eroded further.

Rock cycle



- Weathering & erosion
- Transportation & deposition
- Compaction & cementation
- Heat & pressure
- Brought up to surface

Weathering - process which breaks down rocks to form sediments

Physical - breakage & disintegration
Chemical - decomposition by reaction with water

Rocks react with hydrosphere, atmosphere & biosphere
Lower T and pressure - than where the rocks were formed, so the structures bcome unstable

Physical weathering
- Mechanical break up ⇒ doesn't change mineral make-up
- Creates fragments - "detritus"
- Detrital classification by size: Coarse grained - boulder, pebble
  Medium grained - sand
  Fine-grained - silt, clay

In a fluvial environment, where rivers is the main transportation agent - in the hills we mostly find pebbles, in the plains - sand, and near the ocean coastal - silt and clay (fine grained)

Types of physical weathering –

- Jointing – thermal expansion, tensile stress
  Results in brittle fracture of Rock body

- Frost / Salt / Root wedging
  - water fills the cracks and in v. cold T, ice forms which has greater volume.
  - coastal area – saltwater fills the cracks – water evaporates – salt forms crystals
- Thermal expansion – large diurnal T variation
- Animal activity

Physically weathered rock fragments move by gravity
Large blocks accumulate as talus below a cliff / slope
Smaller fragments are carried away by water / wind

Jointing

Deep crustal rocks are formed at high P and T
As erosion removes material, deep rocks are exposed to the surface where they cool & expand at surface
This causes fractures in the rock called 'Joints' which may exhibit a vareity of geometries

- Igneous plutons crack in onion-like 'exfoliation' layers which break off as sheets that slide off
  Over time, this creates domed remnants

Biological Weathering

Organisms secrete important chemical weathering agents – roots, fungi, lichens, bacteria
Secrete organic acids that attack minerals

Chemical Weathering

- Reaction with water disintegrates many minerals
- Tropical weathering is intensive - turns rocks into heavily decomposed "saprolite"
- It's virtually absent in deserts
- It's hard to say anything about the rate of chemical weathering or how much a rock has weathered by looking at it.
- This process forms stable minerals from unstable precursors
- Common weathering reactions -

1. Dissolution

Halite, gypsum, calcite dissolve

$$CaCO_3 + H_2CO_3 + 2H^+ \longrightarrow 2H_2O + 2CO_2 + Ca^{2+}$$

Acidity (acid rain) enhances this effect.

2. Hydrolysis

Water breaks down cation bonds in silicate minerals and yields -

Dissolved cations + alteration residues (clay minerals) (iron oxide [rust])

$$\underset{\text{Feldspar}}{2KAlSi_3O_8} + 2H^+ + 9H_2O \longrightarrow \underset{\text{Kaolinite}}{Al_2Si_2O_5(OH)_4} + \underset{\text{Silica}}{4H_4SiO_4} + 2K^+$$

Rock is broken into lose grains - feldspar turns into clay (which washes away) and quartz grains become rounded.

Recall : Bowmen's Reaction Series

Olivine (900 - 1100°C) ⟶ Quartz (400°C)

* Least resistant

* Most resistant to weathering

Silicate and Carbonate weathering equations
Basically, CO_2 is brought down

3. Oxidation
Metal loses electrons – rusting

$$2FeS_2 + O_2 \rightarrow Fe_2O_3 + 2S$$

Pyrite — Hematite — dissolved S

4. Hydration
Absorption of water into a mineral structure
⇒ expansion. important process in some clay minerals

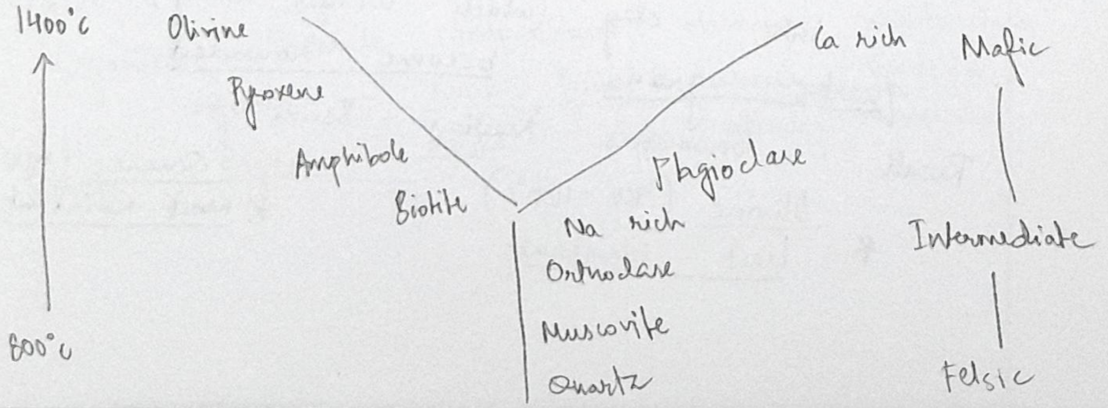
$$2Fe_2O_3 + 3H_2O \rightarrow 2Fe(OH)_3$$

Limonite

Differential weathering – mineral sensitivity
Silicate weathering susceptibility is predicted by stability – ↑T, P minerals weather quickly at surface
↓T, P minerals stable at surface

Weathering rates and controls
Often acts as positive feedback loop –
- ↑ surface area accelerates chemical reaction
- Chemical weathering increases surface area by decomposition

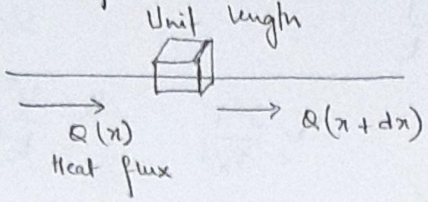
Weathering index – measure of changes in bulk chemistry as weathering progresses
Bulk major element oxides chemistry is used to give a single value to characterize weathering profiles



Lecture ~~N~~ – Discussion

Q.17 of Assignment

Diffusion Equation for wire (1d)



Flux : Amt. of heat per unit area per unit time

$\frac{\Delta H}{\Delta t}$: Amount of heat in unit length of wire in Δt time.

$$\frac{\Delta H}{\Delta t} = -\left[Q(x+dx) - Q(x)\right]$$

Here, C_p is volumetric

$$\Delta H = C_p \Delta T$$

In our case,

$$C_p \frac{\Delta T}{\Delta t} \cdot \Delta x = -\left(Q(x+dx) - Q(x)\right)$$

$$C_p \frac{\Delta T}{\Delta t} = -\frac{\partial Q}{\partial x}$$

$$C_p \frac{\partial T}{\partial t} = -\frac{\partial Q}{\partial x} = \frac{-\partial}{\partial x}\left(-K \frac{\partial T}{\partial x}\right)$$

$$\boxed{Q = -K\frac{\partial T}{\partial x}}$$

$$C_p \frac{\partial T}{\partial t} = +K \frac{\partial^2 T}{\partial x^2}$$

$$\boxed{\frac{\partial T}{\partial t} = \frac{K}{C_p}\frac{\partial^2 T}{\partial x^2}}$$

} Assuming conduction, we can arrive at T profile in space and time

Also used in hillslope processes.

Instead of C_p, thermal diffusivity (κ) is used

Solving differential equation – non-dimensionalise it.

T' : non-dimensional T : $T' = \frac{T}{T_0}$

Similarly, $t' = t/t_0$ $x' = \frac{x}{L}$

So we have —

$$\frac{\partial T'}{\partial t'} = \frac{\partial T'}{\partial T} \cdot \frac{\partial T}{\partial t} \cdot \frac{\partial t}{\partial t'}$$

$$= T_0 \cdot \downarrow \cdot \frac{1}{t_0}$$

$$\therefore \frac{t_0}{T_0} \frac{\partial T'}{\partial t'} = \frac{\partial T}{\partial t} = K \frac{\partial^2 T}{\partial x^2} = K \cdot \frac{t_0}{L^2} \frac{\partial^2 T'}{\partial x'^2}$$

$\frac{t_0}{L^2}$ Making $\to 1$

So, $L = \sqrt{K t_0}$

For a boulder weathering,

time scale : 10^3 s

$K \sim 1-2$ mm²/s $L \sim 20-30$ mm

Length scale of efficient process : the process decays to negligible rates.

After this length, the process decays to negligible rates.

25/4

Lecture 13

Weathering

Physical process — Thermal expansion

Different rocks are made of different minerals which, based on mineral structure, has different expansion coefficients & albedo.

If these is differential expansion, then stress is created and cracks appear.

Thermal stress ΔT : change in T

$V_f - V_i = \alpha V_i \Delta T$ α : coefficient of thermal exp

$$\alpha = \frac{1}{V}\left(\frac{\partial V}{\partial T}\right)_P$$

Formally α is the increase in volume of unit volume block with unit increase in T.

$$\sigma = \frac{E}{1-\nu}\,\varepsilon$$

ε : volumetric strain $\varepsilon = \frac{\delta V}{V}$

σ : stress at boundary

E : Normalised Young's modulus

Conduction through rock –

Recall : $\delta \sim \sqrt{Kt}$ K: diffusivity t: time

δ : length scale

if $K \sim 1\text{-}2$
$t \sim 1000$ s

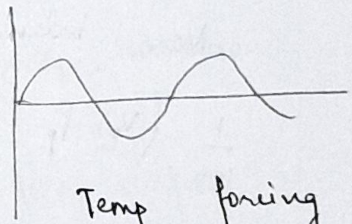
$\delta \sim$ 30 mm – Natural l scale

Temp of mafic > Temp of felsic (quartz, feldspar granite) [by ~2-3°C]

Solving heat conduction equation (assuming boundary condn)
Temp at different depths, times of day

$$T = T_0 + T_{amp}\, e^{-z/z^*} \sin\left(\frac{2\pi}{P}t - \frac{z}{z^*}\right)$$ *

How temp forcing trickles down to various depths of rock


Temp forcing

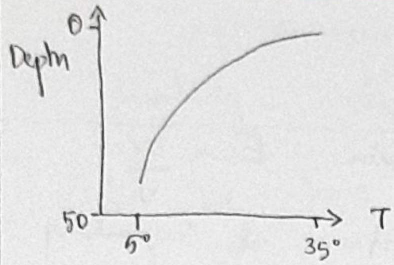
The temp forcing at the surface only affects the rock to a depth of natural lengthscale

Lag of $\frac{z}{z^*}$ – greater the depth, greater the lag
Response is instantaneous at the surface

$$z^* = \sqrt{\frac{KP}{\pi}}$$ P: periodicity

z^* : Natural length scale

Forest fire – 1-2 mm – very short period
Diurnal variation – 1 m
Annual variation – 5-10 m



$z^* = 21.4$ cm

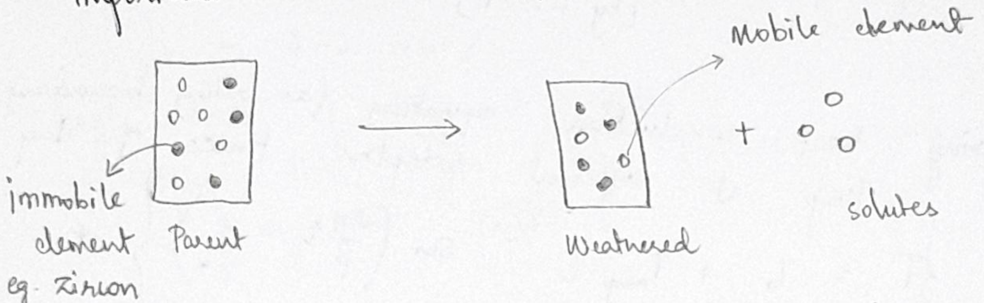
$\alpha = 1.7$ mm²/s

With depth, the effect's amplitude decreases and peak shifts right because of lag

Chemical Weathering

Quantification

Increase in loss of rock mass strength increases denudation in the area. So its important to quantify the rate and keep track.



Mass balance equation

$$\frac{1}{100}\left(V_p, \rho_p, C_{j,p}\right) = \frac{1}{100}\left(V_w \rho_w C_{j,w}\right) - m_{j\cdot flux}$$

(V_p – unknown; parent; weathered; concentration)

$$\frac{1}{100}\left(V_p \rho_p C_{i,p}\right) = \frac{1}{100}\left(V_w \rho_w C_{i,w}\right)$$

— find V_p from this eq^n & substitute to find $m_{j\cdot flux}$

Weathering index as a mass transfer coefficient —

$$\tau_j = \frac{m_{j\cdot flux}}{V_p \rho_p C_{j,p}} \times 100 = \frac{C_{i,p} C_{j,w}}{C_{i,w} C_{j,p}} - 1$$

→ Can be +ve or negative

For unweathered rock, $\tau_j = 0$

soil – -ve values

For weathered rock (b/w saprolite and unweathered rock) T_j is positive ∵ the leached elements from the soil are deposited on this.

Various weathering indices are based on this element.

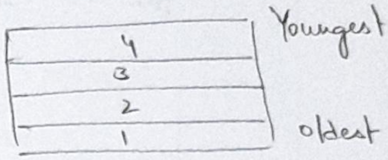
Assumed that Al_2O_3 is an immobile mineral

Discussion session — 26/4

8/5

Lecture 14 — Dates & rates

Time records in landscape



Surface process rate can be calculated by know the age of different layers.

Short time scale ~ 10^3 yr   long scale ~ > 10^6 year

Different evidences & techniques tell us about different scales with varying resolution.

Basics of Relative dating

→ Laws of Stratigraphy — layer at the surface is youngest layer.

If faulting is recent than intrusion, then even the intrusion will be displaced

→ Extent of weathering of a surface. If we have diffusion eqn, you can quantitatively date

→ Roughness of surface : smooth surface ⇒ old
rough surface ⇒ young-ish

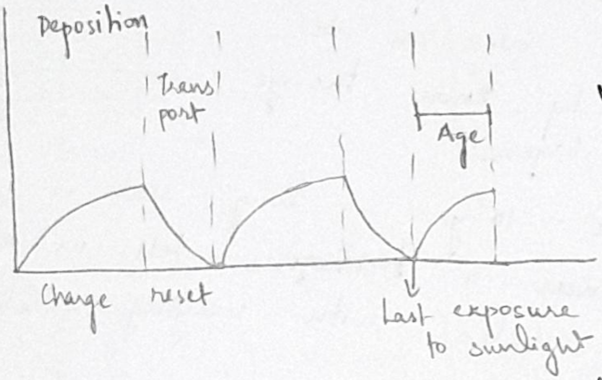
Optically stimulated luminescence techniques.

The luminescence of minerals is used to date the layers.

Different mineral particles get deposited and buried. Longer it stays buried, the more the mineral has been accumulated.

This technique is used to date the last time the quartz sediment was exposed to sunlight & zeroed of any previous luminescence signal.

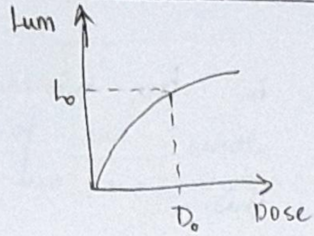
Buried quartz minerals accumulate luminescence signal as ionizing radiation excites electrons in the crystal lattice.



The natural radiation comes from decay of naturally occurring uranium / thorium in the sediments.

The electrons go from valence bands to crystal defect traps in between valence and conduction bands. Longer the sediment stays buried, more electrons are present in these traps. Using them, the age of surface can be determined.

The sample is irradiated with blue light so electrons go to conduction band. As they come back to ground state, they emit natural luminescence. Corresponding to each amt of irradiated dose, the mineral emits some luminescence which is plotted.



Then for the natural luminescence L_0, we can calculate the dose required

$$\frac{D}{D_R} = t$$

Dose rate (D_R) is proportional to decay rate of Ur, Th

When the sample is exposed to sunlight, then all luminescence is lost (?). Also if some remnant dose can exists, it'll make us overestimate the age.

This is more useful for fluvial or glacial sediments than aeolian.

Radiometric dating methods

→ $^{14}_{6}C \longrightarrow {}^{14}_{7}N + e^{\ominus} + \bar{\nu}_e$

$$\boxed{\frac{dN}{dt} = -\lambda N} \Rightarrow \boxed{N = N_0 e^{-\lambda t}}$$

^{14}C Half life ~ 5,500 year

At ~ 10 times $T_{1/2}$, then the parent element is negligible. So limitation : 5 - 6 × $T_{1/2}$.

Drawback

* Non-steady production : cosmogenic rays produce ^{14}C. They get incorporated into plants by photosynthesis. The magnetic feild somehow affects the production rates spatially (which has to be corrected for.
* Reservoir effect — ^{14}C can get sequestered in ocean or atmosphere (not significant residence time). But if ^{14}C is sequestered in ocean, then our estimation of N_0 will be flawed

Other cosmogenic Radionuclides

^{10}Be, ^{26}Al, ^{36}Cl, ^{14}C

Produced by interaction of cosmic rays and earth surface materials.

^{10}Be and ^{26}Al are formed in the mineral in situ, while other atoms are formed in the atmosphere. Using this, we can talk about various earth surfaces

Half life. Be – 1.3 Myr
Al – 0.7 Myrs

They're formed by interaction with common minerals – quartz, silica

Spallogenic method – by primary neutrinos at the surface

Muogenic method – secondary neutrino interaction.

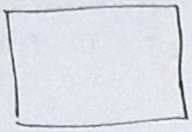
These interactions occur effectively till a certain depth, called length scale of CN interaction – $z^* = \frac{\lambda}{P}$ λ : mean free pathway

$$P_z = P_0 e^{-z/z^*}$$

P : production rate
z : depth
P_0 : production rate at surface

$z^* \sim$ 3-4 m

The production of radionuclides
- decreases with altitude of surface (∵ reduced interaction with atmosphere)
- Latitude (∵ of magnetic fields)
- Topography



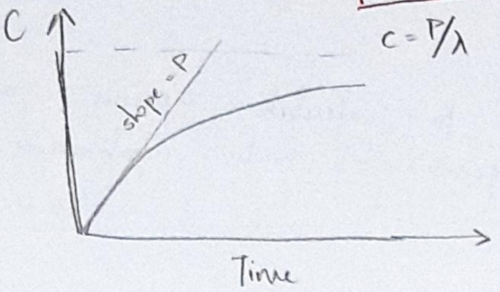
$x \leftarrow dx \rightarrow x+dx$

$$\frac{dC}{dt} = P[1-\lambda C]$$

P: production rate

C: conc of Be / Al

$$C = \frac{P}{\lambda}\left[1-e^{-\lambda t}\right]$$

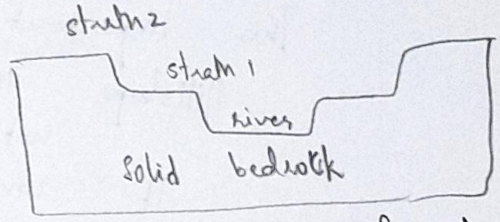


As $t \rightarrow \infty$, $C \rightarrow \frac{P}{\lambda}$

After 6-8 M yrs, the method saturates and its not v. useful.

Upto 70-80 k years, $-\lambda C$ term can be ignored and simple $t = \frac{C}{P}$ can be used to calculate the time.

Using this, the rate of down cutting of rock by river can be estimated



If the surface is made of unconsolidated sediment, the sediment at the terrace aren't being exposed to cosmic rays for 1st time. They'd have been accumulated $^{10}Be/^{26}Al$ when getting deposited. So samples are taken at multiple depths & where the value stabilizes, that's taken as true value.

This can also be used to talk about exhumation rate. To do this, we assume that, at a depth < 4-5 m, conc of ^{10}Be is negligible. As upper layers get eroded, the lower ones are exposed and their ^{10}Be conc. increases. Faster the erosion, faster the exhumation, so production rate at that layer will increase faster.

(34) When ~~surface~~ layer is at surface, it would have collected some ^{10}Be. That is given by $\int P . dt$. The faster its exhumed, the lower total conc.

there, $P = P_0 e^{-\varepsilon t / z^*}$

z^* is material specific

ε; erosion rate

$$C = P_0 \frac{z^*}{\varepsilon}$$

This allows one to calculate erosion rate. at any point — wide application.

9/05

Lec 15

Fluvial geomorphology

Two important components — channels & hillslopes

Hillslope is everything in the landscape that's not linked to channel

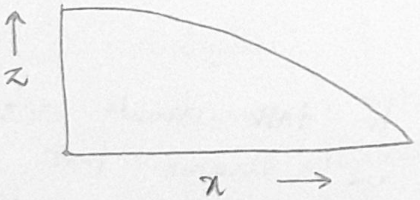
It covers most of area of landscape

Hillslopes consist of mountain ranges, terrace surface

An important feature of hillslope — convex topography

G K Gilbert — characterised the convex topography by a parabolic equation.

$$z = c_1 [1 - c_2 x^2]$$



After differentiating,

$$\frac{\partial^2 z}{\partial x^2} + A = 0$$

curvature; A: const

similar to diffusion eqn — thermal gradient

Just like thermal diffusion gradient,

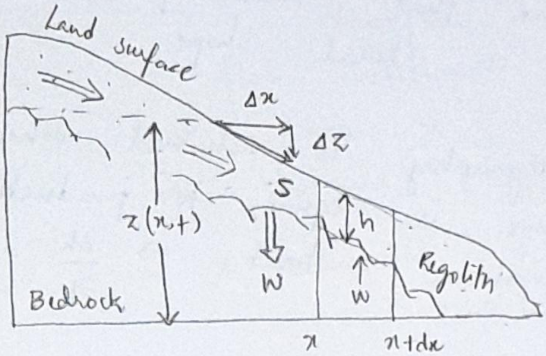
$$\frac{\partial z}{\partial t} = c \frac{\partial^2 z}{\partial x^2} + A$$

c → Hillslope diffusivity

$\frac{\partial^2 z}{\partial x^2}$ → curvature

A → Source of uplift.

Source of topography (elevation) - rock / tectonic uplift



When we talk about hillslope processes, we consider the regolith (soil layer, rock fragments) which lies above the bedrock.

Regolith is generated by physical & chemical weathering and the transportation process is called erosion. Together they came denudation.

Consider a cross-section of regolith

q_x : incoming flux at x

q_{x+dx} : outgoing flux at $x+dx$

w : weathering rate of bedrock i.e production rate of regolith.

These factors influence h -

erosion rate $\dot{w} = \frac{de}{dt}$

ρ_r : density of rock

$$q_x - q_{x+dx} = \frac{\Delta M}{\Delta t} + \rho_r \dot{w}$$

This is the basic conservation eq^n.

bulk ρ ; change in vol when area = 1

$$q_x - q_{x+dx} = \frac{\rho_b \cdot \Delta x \Delta h}{\Delta t} + \rho_r \dot{w}$$

Taking $\Delta h, \Delta t \to 0$

$$-\frac{\partial q}{\partial x} = \rho_b \frac{\partial h}{\partial t} + \rho_r \dot{w}$$

If we assume $q_x \propto -\frac{\partial z}{\partial x}$, we can model this using diffusion eq^n and arrive at steady state convex shape

$$q_x = -k \frac{\partial z}{\partial x}$$

$$\boxed{K \frac{\partial^2 z}{\partial x^2} = \rho_b \frac{\partial h}{\partial t} + \rho_r \dot{w}}$$: Diffusion Eqⁿ

if we assume that local flux is dependent on local slope.

Steady state topography is acheived when net outgoing flux is equal to production rate w i.e h remains constant $\Rightarrow \frac{\partial h}{\partial t} = 0$

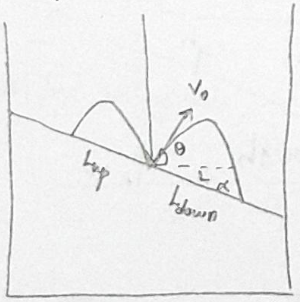
$$\Rightarrow K \frac{\partial^2 z}{\partial x^2} = \rho_r \dot{w}$$

$$\therefore \left\{ \frac{\partial^2 z}{\partial x^2} = \frac{\rho_r}{K} \cdot \dot{w} \right\}$$

If we solve this, we'll get a 2nd order parabolic eqⁿ for hillslope.

Assigning physical meaning to diffusivity compels us to understand different mechanism by which transport of regolith occurs —

1 Rainsplash transport



$$L_{net} = \frac{L_{down} - L_{up}}{2}$$

Main agent that transports sediment is gravity.

Individual grain vs bulk mass movement

Imp: Linear relation of flux with local gradient & conservation of mass

When raindrops impact a hillslope with loose sediment grains, it can jump & travel a distance through ballistic motion.

A raindrop impacts many grains. When we average over the entire rainfall, the net transport of grains in a flat hillslope will be 0.

If we know v_0, we can calculate range of distance travelled by grain.

$$z = x \tan\theta - \frac{g}{2 v_0^2 \cos^2\theta} x^2$$ vertical position wrt horizonal position

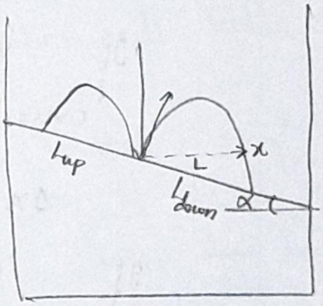
$$L = \frac{2 v_0^2}{g} \cos\theta \sin\theta$$ Maximum displacement

θ : angle of v_0 wrt horizontal

$$dL = \frac{2 v_0^2 \cos^2\theta}{g} \tan\alpha$$ α : inclination of slope

For flat hillslope, $\alpha = 0 \Rightarrow dL = 0$

$$z = -\tan\alpha \cdot x$$



$$L_{down} = \frac{2 v_0^2 \cos^2\theta}{g} (\tan\alpha + \tan\theta)$$

$$L_{net} = \frac{4 v_0^2 \cos^2\theta}{g} \tan\alpha$$

$V_0 \propto d^4$ where d : diameter of rainfall

Larger the rainfall, more the displacement.

Short, intense rainfall is effective in transporting sediment.

$$q_x = n \cdot m N L_{net} \Rightarrow q_x \propto \tan\alpha$$

flux; local gradient

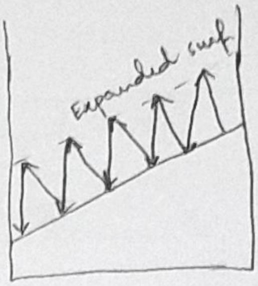
m : mass of raindrop

n : no. of sediment grains

N : no. of raindrops

2. Creep process (See slide 5)

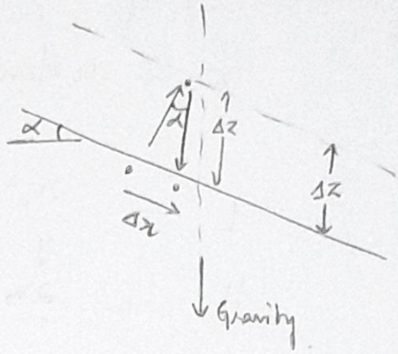
Another process through which grains are transported. This manifests on landscape by bending of poles / trees on the slope.



Mechanism:

The surface (& grain on it) moves (expands & contracts) due to thermal reasons.

The grain moves up due to expansion. When soil contracts, gravity acts on it. So its displaced by Δx



$$\boxed{\Delta x = \Delta z \cdot \sin\alpha}$$

Since α is pretty small, $\sin\alpha \approx \tan\alpha$ (local gradient)

If we just use Δx, net is usually overestimated. So a recovery factor is introduced.

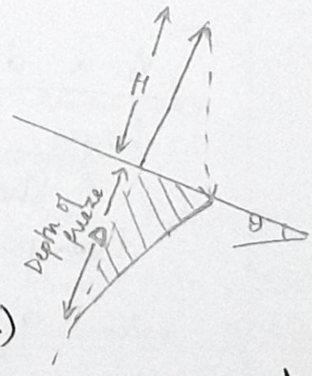
$$\Delta x = \Delta z \cdot \tan\alpha \, (1 - r)$$

If $r = 1$, $\Delta x = 0$ i.e. no displacement

$r = 0$, max displacement.

3. Freeze & thaw process.

One of the processes that affects surface displacement.



$$V = \sum_{i=1}^{n} H_i \tan\theta$$ Summation over all sediments

Annual displacement profile (after n cycles)

$$V(z) = \sum_{i=1}^{n} \frac{D_i - z}{D_i} H_i \tan\theta$$ – displacement of particles below the surface

$$q = \int_0^{D_i} H_i \tan\theta \, \frac{D_i - z}{D_i} \, dz$$: Flux

* for rainsplash & creep. $f(\nabla H) = \nabla H$

\# $\frac{dH}{dt} = -\vec{\nabla} \cdot (-D(\nabla H)) = +D \cdot \frac{\partial^2 H}{\partial x^2}$

$$q_i = \frac{1}{2} \beta D_i^2 \tan\theta \, \rho_b \qquad \beta = \frac{H_i}{D_i}$$

Individual flux

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Lecture 16 _ Hillslope P2

q_s : sediment flux

$\frac{\partial h}{\partial t} = -\vec{\nabla} \cdot q_s$ Cont. of mass eq^n (divergence of flux).

$q_s = -D f(\nabla h)$ Δh : topography gradient (∵ its $\propto \tan\theta$ ie local slope)

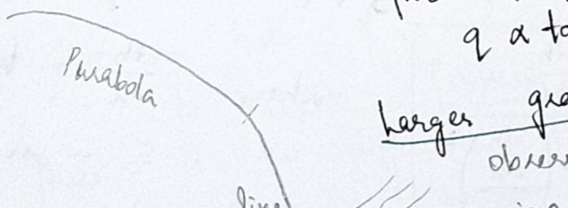
Also, $q_s = -D \nabla h$ *

Linear diffusion eq^n : $\frac{\partial h}{\partial t} = D \frac{\partial^2 h}{\partial t^2}$ \# $+ \; S$ → source term $+ \; U$ → rate of bedrock erosion rate + → surface uplift

If we solve it for steady state landscape ie $\frac{dh}{dt} = 0$, we'll get a parabolic / convex solution. It holds for most physical hillslopes.

Sometimes, the profile of hillslope is better characterised with a parabola & a st. line near the river. This profile could be seen because $q \propto \tan\alpha$ might not hold everywhere

Larger gradient of hillslope generally observed near tectonically active region



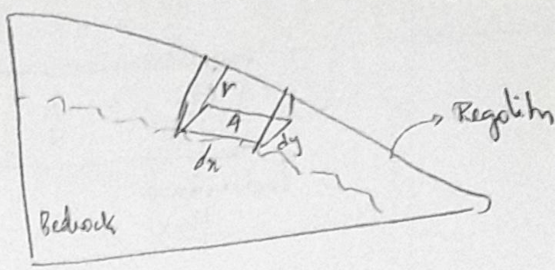
Refer to figure in Pg. 35

Consider a regolith unit. The cont. of mass eq^n will be the same. To find the relation between flux and slope of local gradient, we'll use force balance approach

(40) Volume V, cross sectional area, A

$q_s = \frac{V}{A} \cdot \bar{v}_s$ $\bar{v}_s$: avg velocity of all grains in regolith unit

$A = dx \cdot dy$



External agents : rainsplash

We'll define a power term (P) - work done in transporting unit regolith in unit time

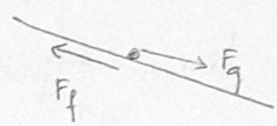
$$\Rightarrow \quad P = F \cdot \bar{v}_s = F \cdot \frac{q_s A}{V}$$

$$\Rightarrow \quad q_s = \frac{V \cdot P}{A F} = \frac{P/A}{F/V}$$

→ Power per unit area

↳ Resisting Force per volume.

$$q_s = \frac{P}{A}\left[\frac{1}{(F/V)_{\text{down-slope}}} - \frac{1}{(F/V)_{\text{upslope}}}\right]$$



For downstream particle, net resisting force : $\underline{F_f - F_g}$

For upstream particle : $\underline{F_f + F_g}$

$F_g = \rho_s \cdot V \cdot g \sin\theta$ $F_f = \mu \rho_s V g \cos\theta$ (relative component of net weight)

V gets cancelled.

$$q_s = \frac{P}{A}\left[\frac{1}{\rho_s g(\sin\theta - \mu\cos\theta)} - \frac{1}{\rho_s g(\sin\theta + \mu\cos\theta)}\right]$$

$$\therefore \quad \boxed{q_s = -D \cdot \frac{\frac{dh}{dx}}{1 - \left(\frac{dh/dx}{S_c}\right)^2}}$$

where $\frac{dh}{dx} = \tan\theta$

$S_c \sim \mu$

$$D = \frac{2P}{A \rho_s g \cdot \mu^2}$$

→ and diffusivity (D)

Flux depends on power of rainfall

If the particle is rolling, then $\mu_{rol} < \mu$

So, diffusivity will be greater

Also note, $D \propto \frac{1}{\rho_s}$ If we reduce bulk density of regolith, it'll increase diffusivity

→ D is spatially homogenous

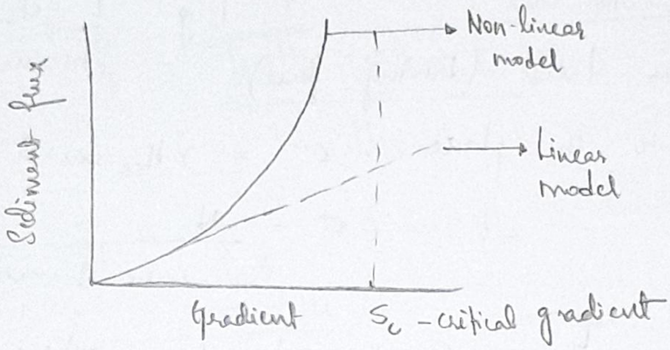
If $S_c \sim 1$ and the slope is less, the denominator goes to 1 approx & we get

$$\boxed{q_s = -D \tan\theta}$$ (same as prev lecture)

where $D = m n N \frac{2v^2 \sin 2\theta}{g}$

This is confirmed by physical observations.

→ Also, if $\frac{dh}{dx} \sim S_c$, then q_s will go to infinity



So, when gradient becomes more steep (straight line profile near the water) the flux increases greatly

Usually $S_c \sim 25-30°$ in mountainous regions
If vegetation is present, the value can be higher
$S_c \in [0.6, 1.4]$

Landslide - bulk of sediment is transported due to mass failure.
There is no diffusive transfer - sediment falls ~~do~~ along a plane of failure

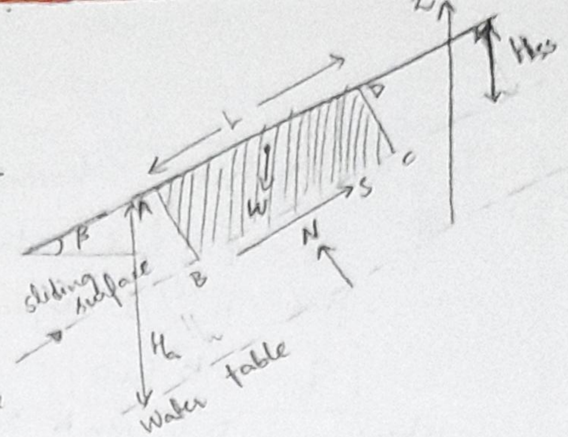
Rock failure - regolith-free hillslope - if rock masses fall down because slope is greater than critical slope

Modelling mass failure - can't use diffusion eqn
we'll use a force ~~mass~~ balance of block of hillslope

(42)

W - weight of hillslope

S : resisting force : of normal to plane component



$S = W \sin\beta$

$N = W \cdot \cos\beta$

$W = \rho_s V \cdot g$

$W = \rho_s \cdot L \cdot H_{ss} \cdot \cos\beta \cdot g$

Height component of sliding surface

$$\tau_{sl} = \gamma H_{ss} \cdot \sin\beta \cdot \cos\beta$$

$\gamma = \rho_s \cdot g$ *

L gets cancelled because of cross-section

Plane parallel shear stress (driving force)

Stress normal to the plane : $\sigma = \gamma H_{ss} \cos^2\beta$

$$\sigma = \frac{N}{\text{Cross-sectional area}}$$

Resisting stress : $\tau_f = C + \sigma \tan\phi$ where C : cohesion

ϕ : critical failure angle $\tan\phi$: μ friction coefficient

Factor of safety : $FS = \frac{\tau_f}{\tau_{sl}}$ $FS = 1$ at time of failure critical value

Slope is stable if $FS > 1$

$\Rightarrow \tau_f = \tau_{sl}$ Consider there's a height that slips when failure occurs. Its given by -

$$H_{ss}^c = \frac{2c}{\gamma \sin 2\beta} \frac{\tan\beta}{(\tan\beta - \tan\phi)}$$

: Critical thickness of slope that can be sustained

The scenario gets complicated when we include the water table i.e. plane of failure is below water table, there's interaction b/w water & sediment. While modelling this, we'll consider its cohesionless.

In this scenario, for cohesionless soil, the factor of safety becomes: $\boxed{FS = \frac{\tan\phi}{\tan\beta}}$ Cohesionless

Failure occurs when $\tan\beta = \tan\phi \Rightarrow \boxed{\beta = \phi}$

ϕ is a property of the material of hillslope

Consider watertable thickness of d (thickness that overlaps with H_{ss})

Bulk density will be lesser than sediment density when water mixes with sediment

$$\rho_b = \frac{\rho_s \cdot (H_{ss} - d) + \rho_w (d)}{H_{ss}}$$ If $d = 0$, $\rho_b = \rho_s$

$\tau_d = \rho_b \cdot g \cdot \sin\beta \cdot \cos\beta \cdot H_{ss}$: Driving force $\gamma_b = \rho_b \cdot g$

$\sigma' = \sigma - u$ → pore pressure – stress due to water present in the pores.

Effective normal stress = differential stress b/w weight of slab & weight of water

$\tau_f = \sigma' \tan(\phi) = (\sigma - u) \tan\phi$

$\sigma = \gamma_b \cdot H_{ss} \cos^2\beta$ $\quad u = \gamma_w \cdot d \cdot \cos^2\beta$

we can simplify this by considering $H_{ss} = d$

If we take $FS = \frac{\tau_f}{\tau_d} = 1$,

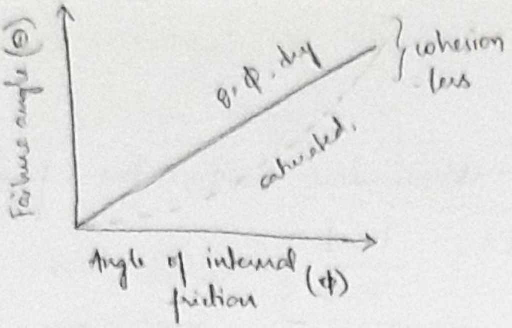
$$\tan\beta = \left(\frac{\rho_b - \rho_w}{\rho_b}\right) \tan\phi$$

$$\boxed{\left(\frac{\rho_b - \rho_w}{\rho_b}\right) \cdot \tan\phi = \tan\beta}$$

⇒ If water is present in hillslope, then angle of failure decreases (by $\boxed{1 - \frac{\rho_w}{\rho_b}}$)

For dry cohesionless sediment, $\beta \sim 30°$

(44)



With increase φ, we can sustain a steeper slope. But for a given φ, angle of failure is lower if water is present.

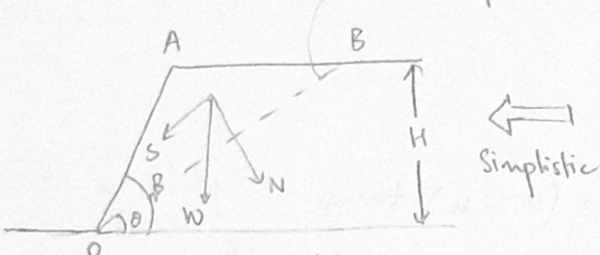
Accordingly, no. of landslides are concentrated during the monsoon months.

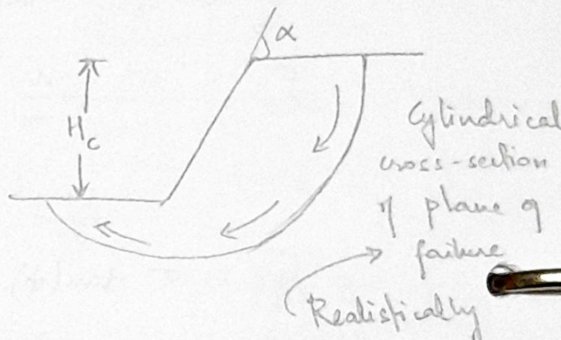
There's also spatial variability – at converged zones, water table is higher, so more local landslides are observed – hillslopes are not infinitely long as we've considered

Having _some_ amount of water in sediment can help attain larger angle of repose because of cohesive property

Realistic landslide

Culmann's model





c′ : cohesion

φ′ : internal angle of friction

$\gamma : \rho . g$

there also, we're trying to calculate H & critical θ

Failure surface assumed to be linear but not parallel to slope surface

Works for high β

$\tau_d = W \sin\theta$: driving stress

$\sigma = W \cos\theta$

$W = \gamma . \frac{1}{2} . AB . H . 1 = \frac{\gamma H^2}{2} (\cot\theta - \cot\beta)$ (1: unit breadth)

$\tau_f = c' + \sigma \tan\phi'$

$\gamma = \rho_b . g$

$FS = \frac{\tau_f}{\tau_d} = 1 \Rightarrow$

Dry hillslope

Shear (driving) stress : $S = W\sin\theta = \frac{\gamma H^2}{2}\sin\theta(\cot\theta - \cot\beta)$

Resisting (normal) ~~stress~~ force : ~~$\tau_f = N\tan\phi$~~

$$N = W\cos\theta = \frac{\gamma H^2}{2}\cos\theta(\cot\theta - \cot\beta)$$

Plane of failure (area) $= \left(\frac{H}{\sin\theta} \times 1\right)$

Dividing weight by area to get stresses –

$$\tau_d = \frac{S}{H/\sin\theta} = \frac{\gamma H}{2}\sin^2\theta(\cot\theta - \cot\beta)$$

$$\sigma = \frac{N}{H/\sin\theta} = \frac{\gamma H}{2}\sin\theta.\cos\theta(\cot\theta - \cot\beta)$$

By Mohr - Coulumb stability criteria,

$$\tau_f = c' + \sigma\tan(\phi)$$

$$c' = \tau_f - \sigma\tan\phi$$

At criticality. $FS = 1 \Rightarrow \tau_f = \tau_d$

$$c' = \frac{\gamma H}{2}\sin^2\theta(\cot\theta - \cot\beta) - \frac{\gamma H}{4}\sin(2\theta)(\cot\theta - \cot\beta).\tan\phi$$

$$c' = \frac{\gamma H}{2}(\cot\theta - \cot\beta)\left[\sin^2\theta - \frac{\sin(2\theta).\tan\phi}{2}\right]$$

∴ Cohesion is only a function of θ (rest all constant for a given slope)

Maximum cohesion : $\frac{\partial c'}{\partial\theta} = 0$ Maximize $c \Rightarrow$ maximize τ_f $\Rightarrow$ maximize H

$$\Rightarrow \frac{\partial}{\partial\theta}\left[\frac{\gamma H}{4}(\cot\theta - \cot\beta)(2\sin^2\theta - \sin 2\theta.\tan\phi)\right] = 0$$

Maximum cohesion → $\boxed{\theta_{crit} = \frac{\beta + \phi}{2}}$ $\boxed{c'_{max} = \frac{\gamma H}{4}\frac{1 - \cos(\beta - \phi)}{\sin\beta\cos\phi}}$

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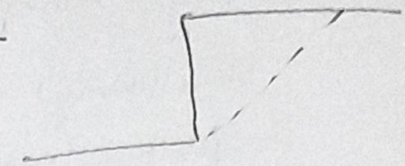
This is at criticality, so

$$H_{max} = \frac{4c'}{\gamma} \frac{\sin\beta \cdot \cos\phi}{1 - \cos(\beta - \phi)}$$

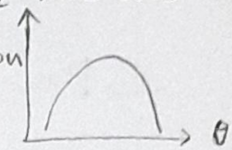
Consider the case of vertical slope –



$\beta = 90°$ $\Rightarrow$ $H_{max} = \frac{4c'}{\gamma} \frac{\cos\phi}{1 - \sin\phi}$

$\Rightarrow$ Without cohesion, no height can be sustained

Remark: $\theta = \frac{\beta + \phi}{2}$ also explains why angle increases with cohesion going upto a point.



Given a cohesion, continuously increasing or decreasing it will initially have higher θ then dip beyond c'_{max}

This explains why weakly watered sediment has more θ. Dry sand has ~~θ < θmax~~ $C < C_{max}$ & as it gets watered, it hits C_{max} and correspondingly θ_{max}

Further water addition reduces c, & hence θ

This in turn reduces H_{max}

After the layers fail from the hillslope, then given some conducive conditions, failed sediments move downslope through a certain path.

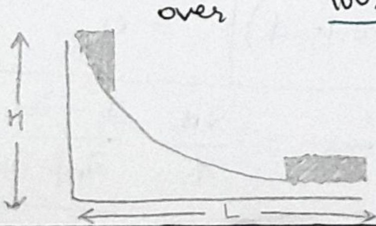
Failure happens along plane of failure – usually curved

Transport of failed materials occurs along a 'run-out' path

This transported material is of 2 types –

a) Debris flow – for mixes of water & sediment
b) Landslide runout – dry slopes or rock failure, no fluid flow

Landslide moves failed material very fast – often over 100s of meters, even for a small one



Material begins from height H & travels upto length L

This $\frac{H}{L}$ ratio can be upto $\left(\frac{1}{20}\right)$.

Highly stochastic flow - define scaling relation H/L instead that can be used to predict a range of dispersion

Can also estimate L numerically

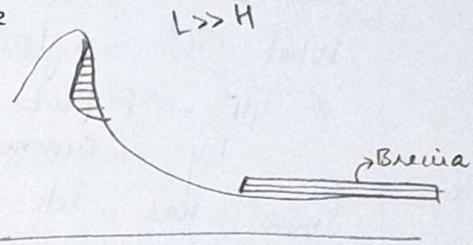
For a landslide runout (also called Long Runout Slides), L >> H. Rock failure has long runout slide to form layer deep below

Interestingly, stratigraphy remains intact ⇒ date relative to layers is maintained

Forms a corrugated ~~lay~~ surface called "Breccia"

Basic eq^n : $\sigma = (\rho_s - \rho_f) V_s g h \cos\theta - P_w$

Use Mohrs - Coulum criterion to derive



σ : Normal stress (resists flow)

ρ_s : density of solid phase

ρ_f : density of fluid phase

V_s : volumetric grain conc

P_w : fluid pressure

We can find P_w - it dictates σ. The rest are constant for a flow. This σ can be used to estimate length scale.

Pore fluid pushes the mass forward, slows down as its conc. reduces by diffusion.

Use continuity of mass eq^n :

$$\frac{\partial P_w}{\partial t} = -\nabla \cdot (q_w)$$

P_w : pore fluid - hydrostatic pressure

q_w : flux of fluid.

$$q_w = \underbrace{\left(-\frac{KE}{\mu}\right)}_{\text{Diffusivity}} (\nabla P_w)$$

K : permeability of rock

E : total estimated elastic and shear bulk modulus

μ : viscosity

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$$\frac{\partial P_w}{\partial t} = \frac{KE}{\mu} \frac{\partial^2 P_w}{\partial x^2} \quad \text{(in 1D)}$$

In a diffusion eqⁿ, we can define a characteristic timescale for a diffusive lengthscale

$$D = \frac{KE}{\mu} \Rightarrow \boxed{t_{diffusion} = \frac{h^2 \mu}{K \cdot E}}$$

Length scale: thickness of debris scale over which fluid diffuses

$$t = \frac{\langle h^2 \rangle}{D}$$

$t_{diffusion} \sim$ 10³ - 10⁵ sec (few min to few hrs)

Typical $v_{flow} \sim$ 1 - 10 m/s

We can estimate distance as, 1 - 10 km

Consistent with feild data!

14/6

Lecture 16 - Glacial geomorphology

What is a glacier?

GIF - F. Paul - Karakoram. where glaciers are fed by summer snow.

Snow has lots of air packets (could be 100-200 g/L) and ice is more dense (~900 g/L). After years of snowfall, the snow gets compacted and forms ice.

This ice creeps - this is because this solid is close to its mp. and it "flows" like a very viscous liquid (same reason why mantle flows).

Mantle - 10^{21} | Water - 10^{-3} | Air - 10^{-5} } Viscosity

Ice - 10^{12}

The snow in mountaintop cannot melt. When they come to lower elevation, they start melting.

The thickness of glacier decreases till it becomes 0 where it melts and forms a stream which feed into larger rivers. ⇒ Glaciers important source of water.

There are many fluctuations in snow cover (even daily) but glaciers only respond to decadal ΔT.

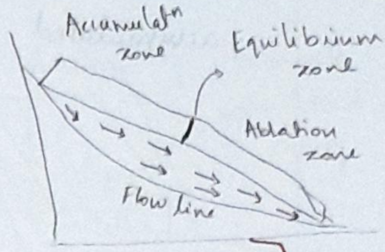
Flow of glaciers ~ 100 m/s width - ~100 m

↳ Acts as a conveyer belt

The glacier, along with ice, contains sediments due to erosion.

Glaciers are a nonequilibrium open system -

- accumulation (snow doesn't melt)
- flow
- ablation (melting)



Its in a steady state if

Tot. accumulation = Total ablation = flux at ELA

So when a glacier is longer than before, it means ablation is more - there's net loss.

If snowfall reduces or T increases, ELA will move up

ELA is basically the line where T = 0.

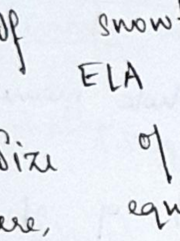
Size of glacier tracks ELA variation

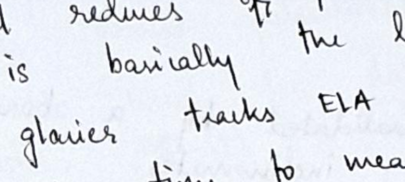
Here, equilibration time to measure T of ELA is decades

So the thermometer is calibrated ~~in through~~ with a complicated formula derived from readings from last ~400 years (Global T construction).

ELA also responds to changing P.

There's some sliding at the interface of ice and bedrock. This is hard to parametrize because its dependent on water pressure, sediments etc.

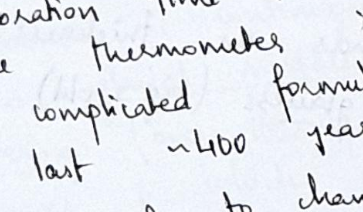




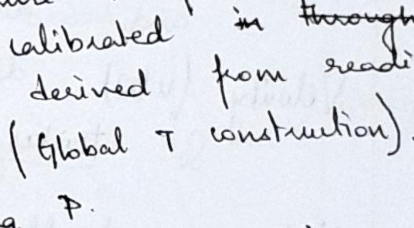
b: local mass balance

Q: ice discharge

$$Q(x) = \int_0^x W(x) \cdot b(x) \cdot dx$$



integral →

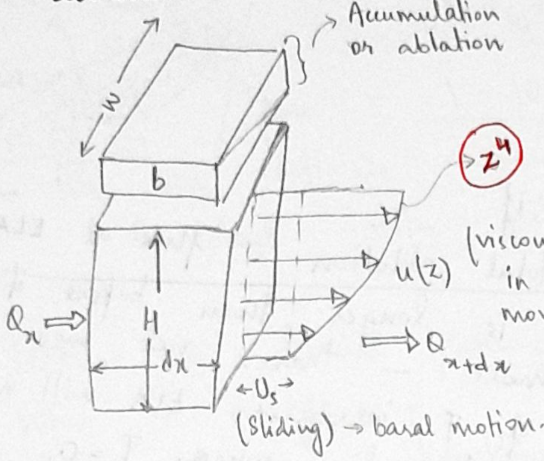


In the ablation zone, the absolute value of decrease increases with decrease in height from ELA

More ice is lost along the glacier?

The ice melt shows a linear curve since T decrease is also linear in that region.

To maintain steady state, at any point, the flux should be able to remove all the accumulated ice upstream of it.



Mass balance model

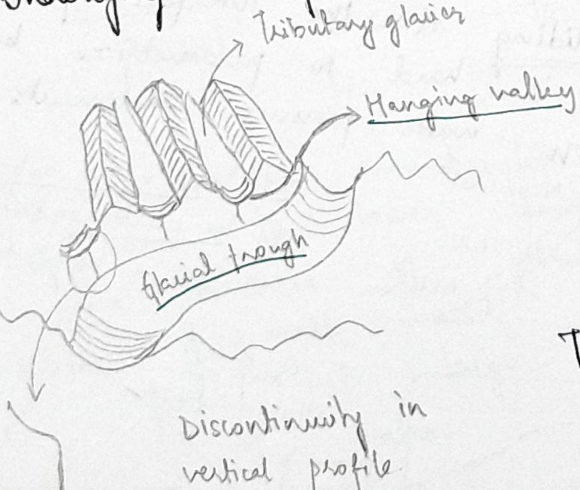
For water, the curve is a parabola (z^2). Here, its much steeper, not near the surface though

As we go down, the stresses increase, so the ice is "thinned out" i.e. internal deformation and shearing

This can be validated by a bore hole measurement and an inclinometer. $u \propto z^4$

Velocity (u(z)) depends on thickness (H) of glacier. So thicker glaciers (big & old) have faster flow.

Underlying landform



This is a schematic of underlying landform of glacier.

Glacial valleys – U shaped

Fluvial valleys – V shaped.

This landform is also dotted with lots of lakes – Pater Noster lakes.

No discontinuity in fluvial

~~Rig~~ Ridges – Arete

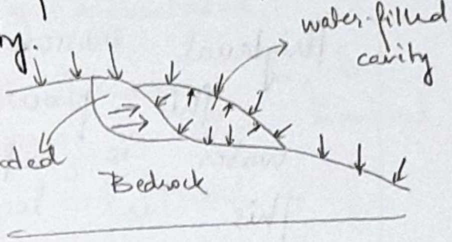
Where ridges meet – Horn → Mt. Everest

How are these landforms formed?

Eroding bedrock – ice sliding over bedrock – scratch it. So the landscape is built – over thousands of years – one scratch at a time.

Technically, they're called **striations**. Bigger ones are called **grooves**.

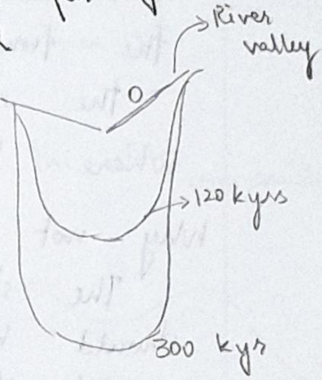
Ice flow can also chip away blocks of bedrock which have a disconinuity. The block gets eroded. Its on the lee side.



TP : **Streamline** – tells us about the past extents of the glacier. Its a part of slope adjacent to glacier, that's very smooth. Tells us about paleoclimate.

Typically. Be is used for dating. But for glacial deposits, OSL (?) is also used.

Glacial bedrock is continuously eroded. This means **glacial surface is also lowering** (over 100s of kyrs).



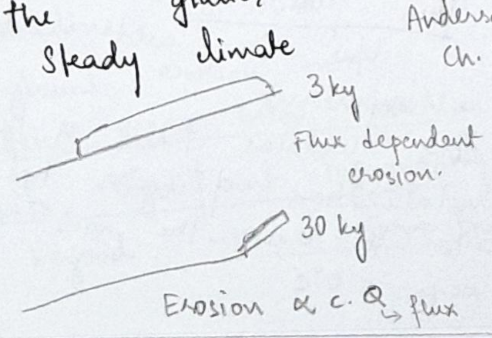
This increases T which creates a (?) negative feedback – over **100s of kilo yrs, glaciers kill themselves**.

Uplift is a process (due to isostasy or tectonic activity) that acts in opposite directions and sustains the glacier.

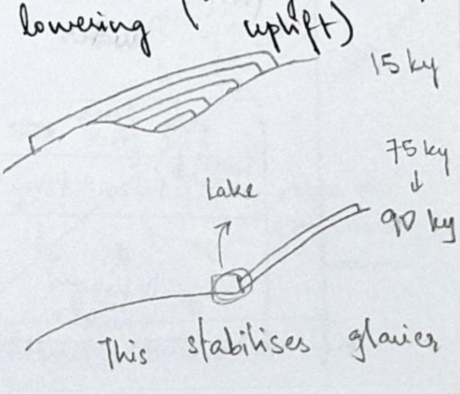
* Steady climate

Anderson Ch. 8

Erosion makes glacier smaller



* ELA lowering (cooling climate/ uplift)



Lowering ELA erodes bedrock in such a way that there's a dip in the bedrock, which after deglaciation forms a lake.

At the junction where a <u>tributary glacier</u> meets the main one, there's a jump in flux. This explains the jump in erosion rates.

⇒ There's a <u>hanging valleys</u> Erosion rate ∝ flux

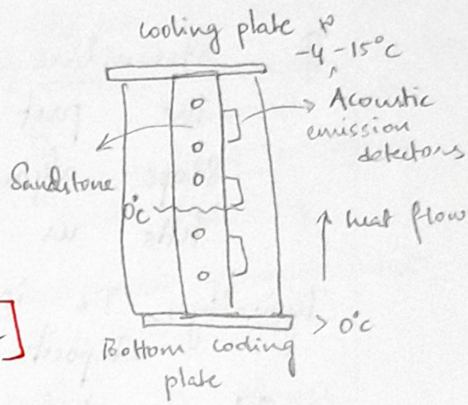
14/6

Lecture 17 — Glacial 02.

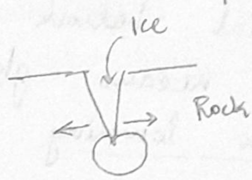
Periglacial erosion — <u>Frost cracking</u> (Ch. 7)

This process breaks up rock in a region where water is present and the right T window

This is localized to certain latitude and altitude ranges.



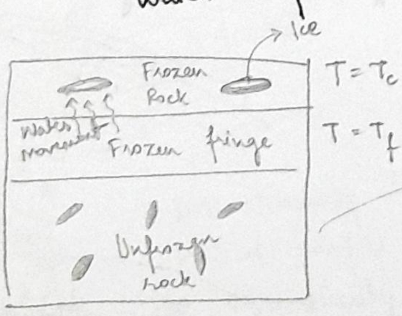
Sound detectors are used to detect frost cracking as the bar of sandstone cracks. Since we know the temperature, we can calculate the range — [−3°C to −6°C] where frost cracking occurs.



Why not at 0°C?

The stress caused by ice should be more than breaking point of the material. <s>ie</s> <u>P inside the crack</u> is greater than P_{atm} ⇒ melting pt of water is lowered

Thus, −3°C to −6°C is the range at which water freezes and (crystallizes) breaks open the crack.



<u>Thin film effects</u> — molecular interaction b/w ice and rock becomes important

Rock contains traces of water. Crystallization of water has to occur at $P > P_0$ to crack the rock ⇒ its m.p. < 0°C

Brozovic et al 1997 , Montgomery et al 2001

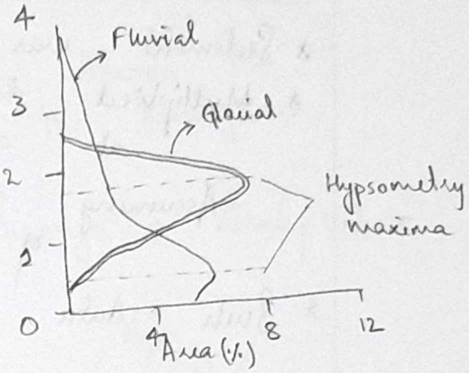
→ Glacial buzz-saw hypothesis

⇒ Glacial effects limiting mountain height.

When a mountain range grows above snow line, it accumulates snow and ice ⇒ erosion rates (glacial & periglacial) increase and this acts as a negative feedback which decreases the height of mountains.

Hence, the local snowline acts as a soft upper limit to the height of mountains.

As we go poleward, snowline decreases, so the height of mountain ranges is also limited.

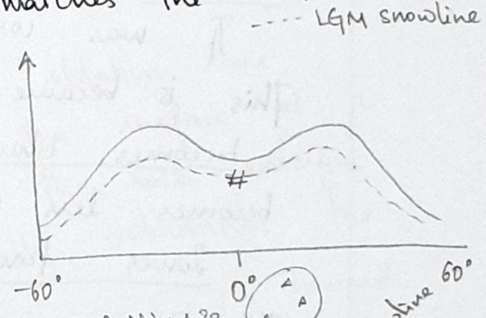


→ Fluvial : bottom-heavy hypsometry curve

Glacial : around glaciers, there's a lot of flat land. So, in a region where glacial erosion dominates, the hypsometry maxima goes up.

buzz saw

The hypsometry maxima almost matches the ELA of the latitude

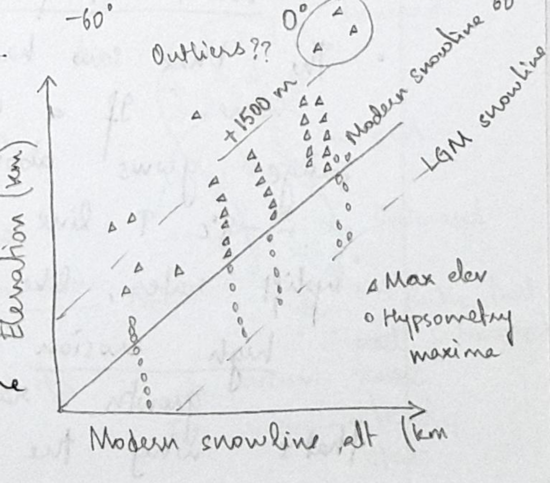


\# ELA is lesser near equator because of constant rainfall

Most hypsometrey max. are between LGM & modern snowline.

Elevation max. are b/w modern snowline & + 1500 m.

∴ Peri-glacial erosion —
- Flattens areas below snowline
- limits significant topography beyond ~1500 m above snowline



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Recall, we said that max. elevation of a mountain on earth is ~9km. Based on this, we can say that this can occur only in certain tropical latitudes.

Is Himalaya-Karakoram an exception? (Outliers in the graph)

Study: Banerjee & Wani, 2018

Satopanth glacier — debris covered glacier — gives rise to Alaknanda. [Bhagirathi & Mandakini also arise from same region].

Glacier floor: 4500 m   Peak of mountain: 7000 m (rock face)

* Sediment was dug to measure debris depth
* Multiplied by rate of flow, we get debris flux at any point

Assuming a steady state, we get an erosion rate of the rock face.

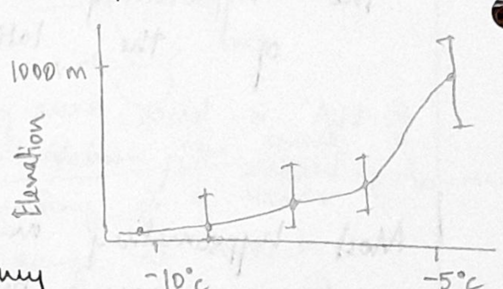
* Such data was collected/collated for 4 glaciers.

Results

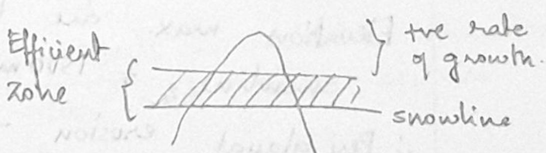
- Variation in erosion rate: from 40m/Ma to ~1km/Ma

It was correlated to Mean temperature

This is because the 'buzz-saw' becomes blunt ie erosion becomes less efficient at lower temperatures.



• The buzz saw has a T efficiency zone. If a mountain range grows above the ~-8°C T line (due to high uplift rates, like in Himalayas), then it escapes the high erosion rates and has a positive growth rate.

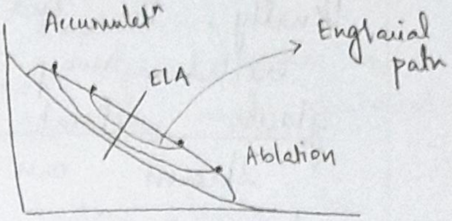


That's why the outliers are from the Himalayan range.

Permafrost – frozen soil – also important factor for shaping geomorphology – read yourself. (Anderson).

Deposition

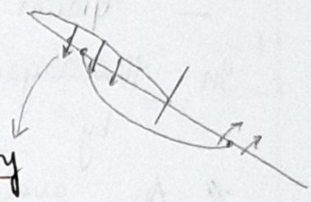
What happens to a rock that falls onto the glacier in accumulation zone?



When its in accumulation zone, it gets buried (under snow) as it moves and once its crosses the ELA, it comes out with the melting ice.

The higher the rock starts, the deeper the englacial path, & later the emergence.

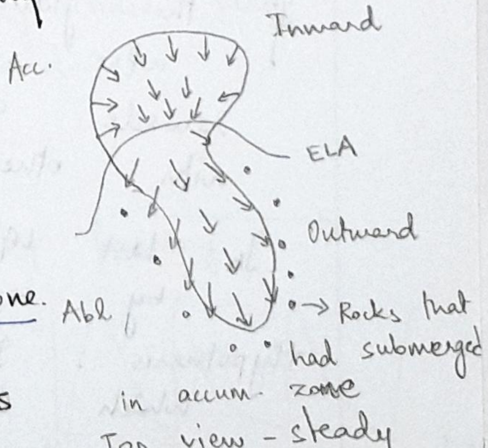
This trajectory can be understood through a steady-state argument.



To maintain steady state, the accumulated snow has to have some downward flux – submergence velocity

This pushes the particle inward

Similarly, as it loses snow in ablation zone, the layers below come to the surface at a rate of emergence flux – the rock emerges with it. This basically determines the movement of the rock.



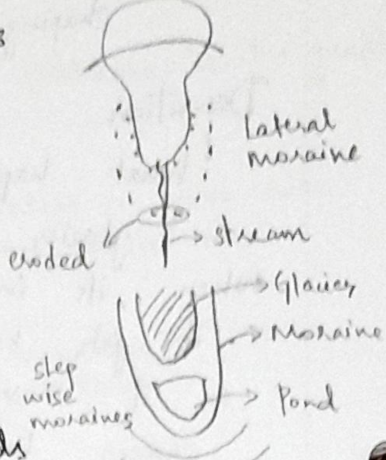
Top view – steady state

From this steady state consideration, we can draw this flux – inward fluxes in the Acc. zone and outward in the ablation zone.

This is when the glacier has steady flux – i.e. maintains its shape for many years.

The rocks submerged come out at the periphery of ablation zone ∵ outward flux. So the glacier deposits debris at its outline.

When glacier shrinks (ELA moves up), then this line of debris remains & it's an indication of previous steady state. Its called a Moraine



Usually, the end moraines are eroded away v. fast. But if glacier retreat is rapid, then the stream can get ponded in the moraine complex. These ponds are very dangerous — they're at high altitude and followed by a steep slope. Moreover the dam is not stable. So if the pond bursts, it comes down with high velocity — Glacial lake Outburst Flood (GLOF).

In Himalayan region, glacial lakes have increased by 50% in last 30 years.

An avalanche can also trigger the GLOF.

Dating the moraines allows us to measure the rate of retreat — glaciers act as self-recording thermometers.

If the glacier advances, all the moraines that were there, they'll get erased. So moraine records are incomplete — has to be bolstered with other paleoclimate records.

In ~~last~~ LGM, glaciers around the world advanced by ~ 10s of km, but not Himalayan glaciers.

Hypothesis : In LGM, monsoon was weakened which could have cut off the moisture / snow supply to the Himalayan glaciers.

Pictures – ~41:00 mins

In Himalaya, moraines are not always well-preserved

Glaciers – solid flowing ⇒ it can carry rocks of any size/mass. Boulders as big as large rooms

P Erratics – when glaciers retreat, they leave the boulders behind called erratics.

These rocks are v. different from rocks in natural surroundings. In size and material.

Deposition of sediment at the bottom leads to subglacial structures such as flutes, drumlins, crag n trail, eskers* (streams enter the ground through ~~there~~ cracks & crevices), and lineations.

P

Refer to pictures in slides

Ice : non-linear viscous solid ⇒ you can see cracks on the surface of the glaciers

Bottom layers ~ like a liquid ; Surface ~ solid

⇒ viscosity increases rapidly as you come to the surface.

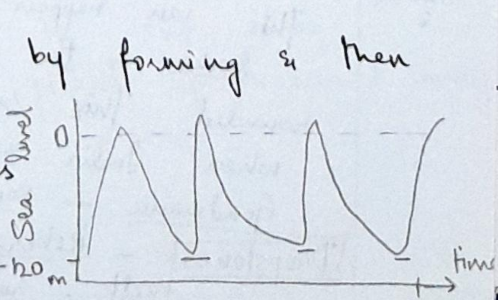
* Crevices give rise to underground drainage channels for streams. When they come out of the surface, they form linear formations called Eskers. – P ~ 46:00 min

Map of Eskers – shows ice sheet of Europe and how its been retreating since last ice age. Eskers have been dated.

Glacial - Interglacial cycle

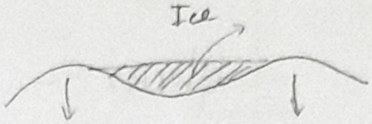
Glaciers play an imp. role by forming & then melting away.

In glacial phases, the sea level decreases. The sea water is stored in glaciers.



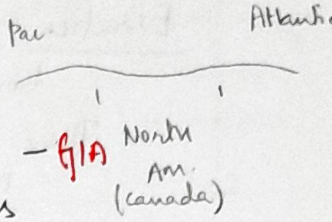
When the glacial ice melted in the inter-glacial phase, due to isostacy, the crust rebounds. The time taken is dependent on viscosity of mantle (10^{24})

So even though LGM was 18,000 yrs ago, its effects are still seen today.



The radial effects can be measured.

The deppressed part is rebounding by ~18 mm/yr whereas the 'bumps' are coming down by ~few mm/yr.

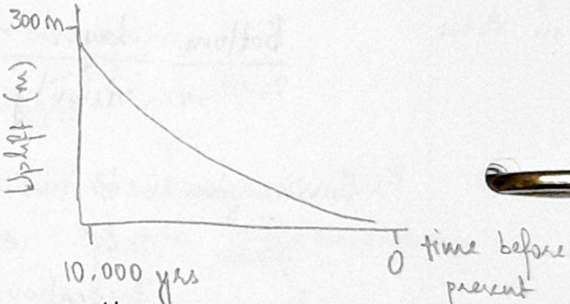


This is called Glacial Isostatic Adjustment – GIA

by loading & unloading, the glaciers affect the shape of earth & local sea level

Because of GIA, the sea level (local) is actually decreasing, not decreasing.

Eg: Series of beaches on Eastern Gotland, Sweden.
This island is uplifting, so local sea level is decreasing, leaving behind beaches.



By dating these beaches, we can see that uplift rate is exponentially decreasing

This curve led to the first proper measurement of the viscosity of the mantle

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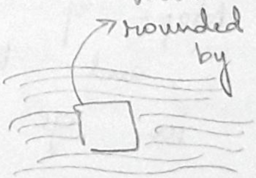
Similar to an Erratic

Rock 3-4 cm Angular

A boulder in middle of sedimentary rock.
How'd it get there?



Boulder in middle of sedimentary stone from Talcher stone belt near (east coast of India) Orissa coast

This can happen when glaciers carry rocks – the edges don't get rounded. This sediment occured when India was a part of Gondwana – south pole saw glaciation.

Dropstones – icebergs carry rocks inside them & when they melt, rocks get deposited in sea. Happens when ice-sheet destabilizes.

Aeolian landforms

Winds are agents of geomorphic evolution.

Landscapes – ripples & dunes

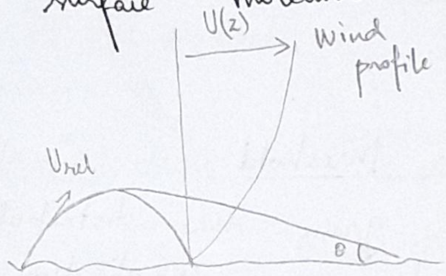
Sediments transferred by wind is applicable to (similar) fluvial transport

Ripples ~ few cm undulating patterns (wind + water)

Consider sand grains –

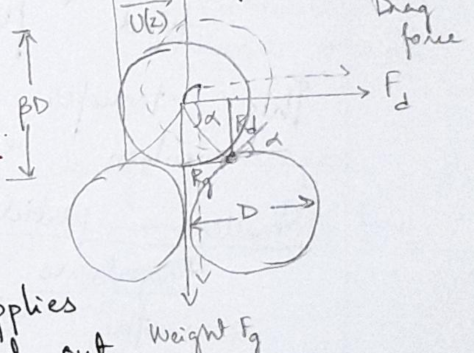
- they're larger that silt / clay particles – they have cohesive interactions also.
- Not all grains are spherical as seen in the model ↙ monodispersed grains

Fluid movement (wind / water) tries to pull the surface molecules along with its motion. due to drag



when wind is flowing along the direction of throw, the particle is carried further and it hits the surface, at a lower angle

A grain at the surface moves if drag force > rolling friction.

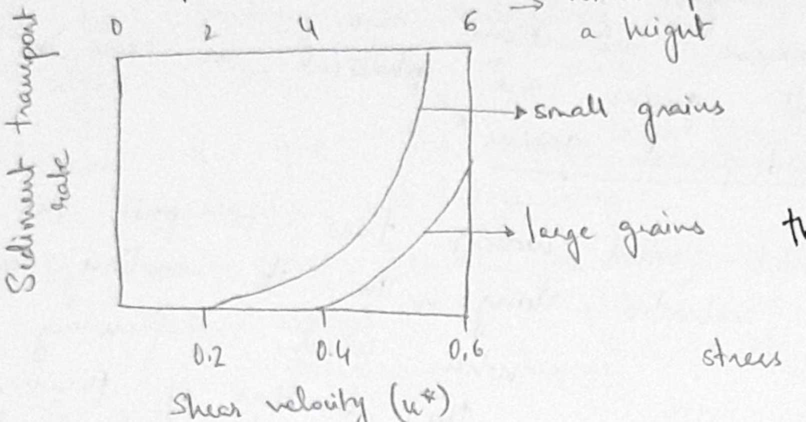


The particle is partly sheilded by residing in the dip. If F_d is strong enough, it applies torque so particle is pushed out of the grove and when its out, drag force increases and its picked up & moved away.

~ Entrainment of particle

Wind speed profile is smooth ∵ avg values are plotted
There are always perturbations / fluctuations greater than avg value which could be below the drag force threshold

When grain particle strikes the surface, it has a lot of momentum – some frictional heat is generated but rest is imparted to surrounding molecules which 'jump' like the initial particle.
This keeps a steady supply of particles to the fluid so it can be transported



Shear velocity is a measure of shear stress at the boundary of the grains.

stress $\propto \rho_f \cdot u_*^2$
↳ density of fluid

Smaller grains – lesser threshold $\propto (u^* - u^*_t)^3$

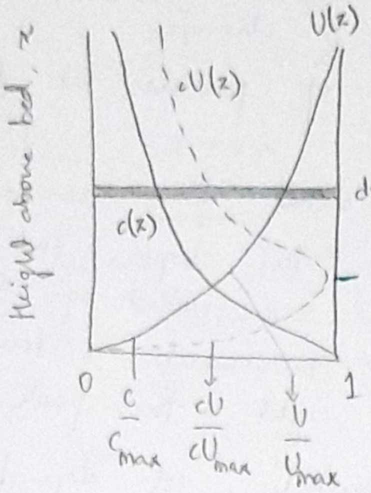
The packing is varied, grain size distribution, wind flow turbulence – complications to model

This transfer of molecules at a constant rate, when equilibrium is acheived

Smaller particles can have more complicated trajectories ∵ they're susceptible to turbulence in the wind. They're likely to stay in air longer & travel longer distances.

This simple transport of sand grains through 'jumps' is known as 'saltation'.

Assume a steady state of particle motion.



c : Sediment conc

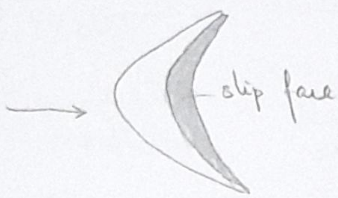
cU : mass flux   U : flow velocity

We get a maxima at a certain height above surface where flux is maximum.

At surface, velocity is low
at great height, sand load is less

Sand dunes

They are dynamic

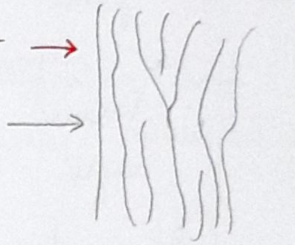


BARCHAN DUNE

Slip face has steeper slope

They appear in sand-limited landscape

If there is abundant sand, then transverse dunes are formed →



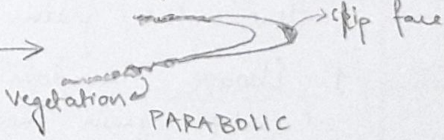
TRANSVERSE

Movement 100m/yr

Star shaped – wind changes direction
multiple slip faces

Linear dunes – 2 slip faces

If there is vegetation, it can change lanscape and shape of dune →



PARABOLIC

Classification of dunes IP

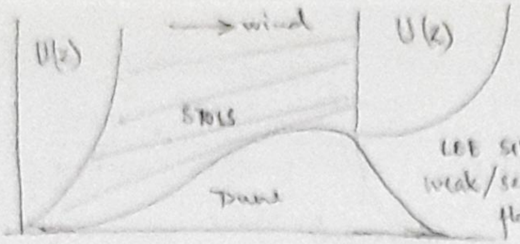
Barchan – crescent, 1 SF

Transverse – asymmetrical ridge, 1SF

Linear – symmetrical ridge, 2 SF

Star – central peak with ≥ 3 SF

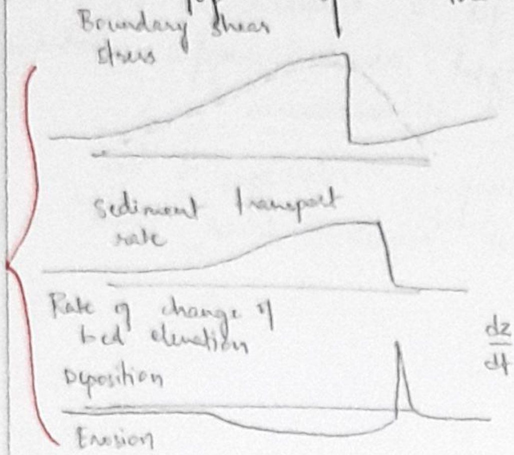
Dome – circular or elliptical mound
No slipface



As the wind flows over the dune, it gets compressed

Wind velocity profile also changes

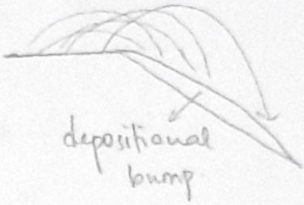
So, there's faster movement of particles at the top of the dune



The wind flow increases along the slope, the particles are picked up and net movement is towards the right, the peak.

The particles get deposited at the peak

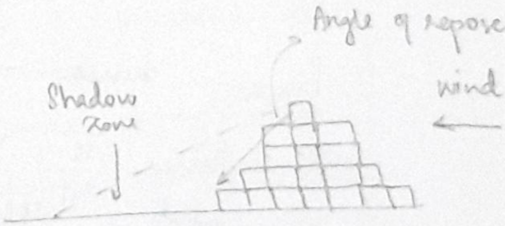
The "peak"



At the peak, when deposited grains overflow (avalanche), a depositional bump. This happens because grains align at a steeper angle than frictional slope

In this way, the dunes must be moving. As high as ~ 100 m/y.

Warner's model



Here, the grains are scattered

1. Choose random site, pick up grain there
2. move it 'l' steps down wind (saltatory length)
3. Put it down with prob. p ⇒ move it further with ~ (1-p)
4. If in shadow zone (cast at 15°), put it down with p=1.

With this model, different kinds of sand dunes can be obtained, by varying amount of sand and direction of wind.

This model can't talk about ~~dune~~ distance b/w dunes (which depends on grain size, wind speed etc). Works qualitatively.

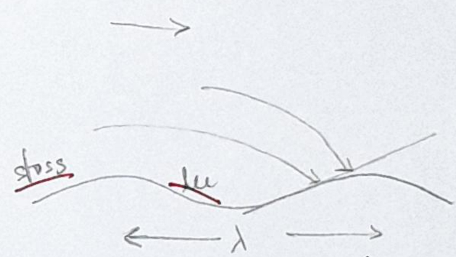
It doesn't take into account interaction b/w dune topography and wind direction.

The things not incorporated just affect details ∴ model predicts the shape correctly.

Simple models work!

Ripples — caused by wind or rivers. They are of ~10 cm scale. This is due to saltation length and so on. (?)

Assume a flat surface in the beginning and there are perturbations, particularly sinusoidal ones of a given wavelength.



Initially, the entrained particles hit the slope facing the wind, rather than other side. So particles from there are likely to be saltated.

stoss side ←

⇒ There's an asymmetry.

The distance travelled by the saltatory particle will be fixed — 'a'. (based on fluid speed) HOP LENGTH

Since they again asymmetrically hit & deposit on one slope, together, λ_0 and a, together will "pick" a new λ of surface. This is the wavelength of ripples. $\boxed{\lambda = 4a}$

The unstable perturbations die out.