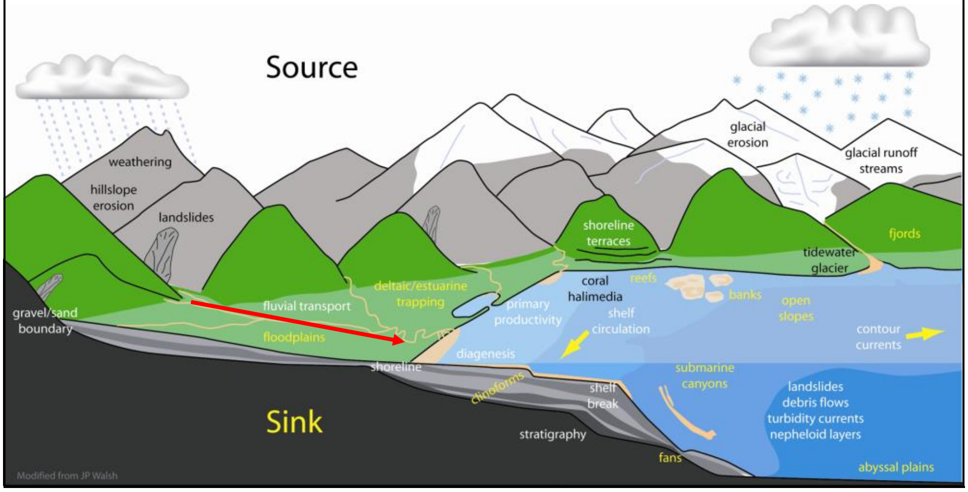
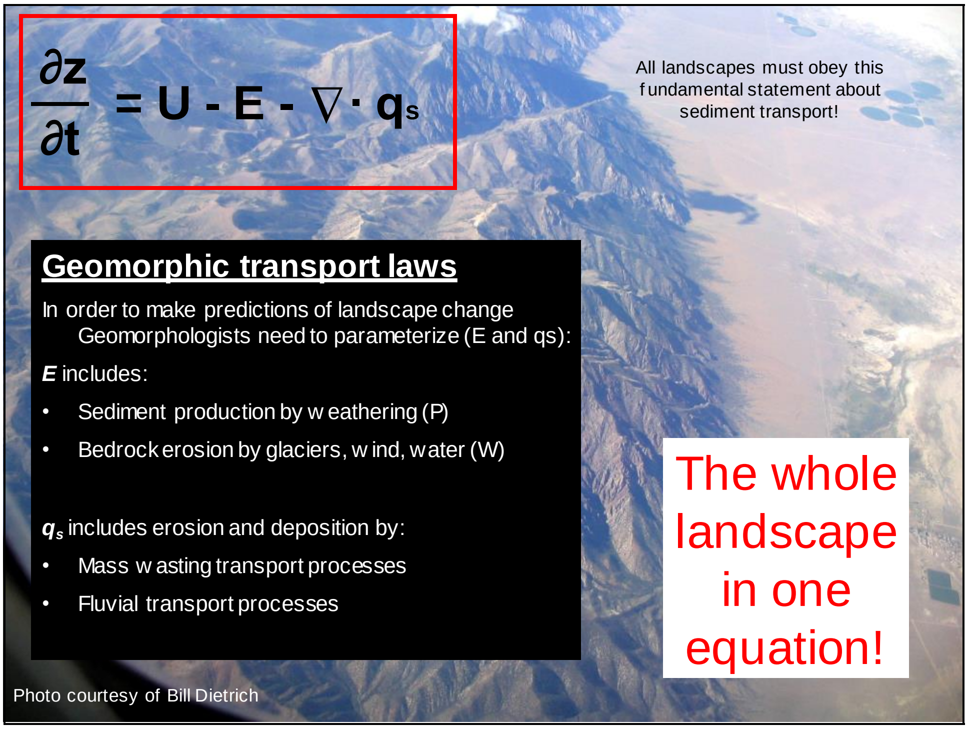
Goals of Rivers Lectures

1. Transition from hillslopes to river channels
    - i) Drainage basins and river networks
    - ii) Long valley profile and downstream fining
2. River channel morphology
3. River processes
    - i) Flow in rivers
    - ii) Sediment transport in rivers
    - iii) Bedrock incision by rivers

How do landscape materials get from valley floors to their ultimate sink (oceans or lakes)?

Sediments are transported from source areas via river channels to the sea. Collectively these processes are called fluvial processes.





$$\frac{\partial z}{\partial t} = U - E - \nabla \cdot q_s$$

All landscapes must obey this fundamental statement about sediment transport!

Geomorphic transport laws

In order to make predictions of landscape change Geomorphologists need to parameterize (E and qs):

E includes:

- Sediment production by weathering (P)
- Bedrock erosion by glaciers, wind, water (W)

q_s includes erosion and deposition by:

- Mass wasting transport processes
- Fluvial transport processes

The whole landscape in one equation!

Photo courtesy of Bill Dietrich

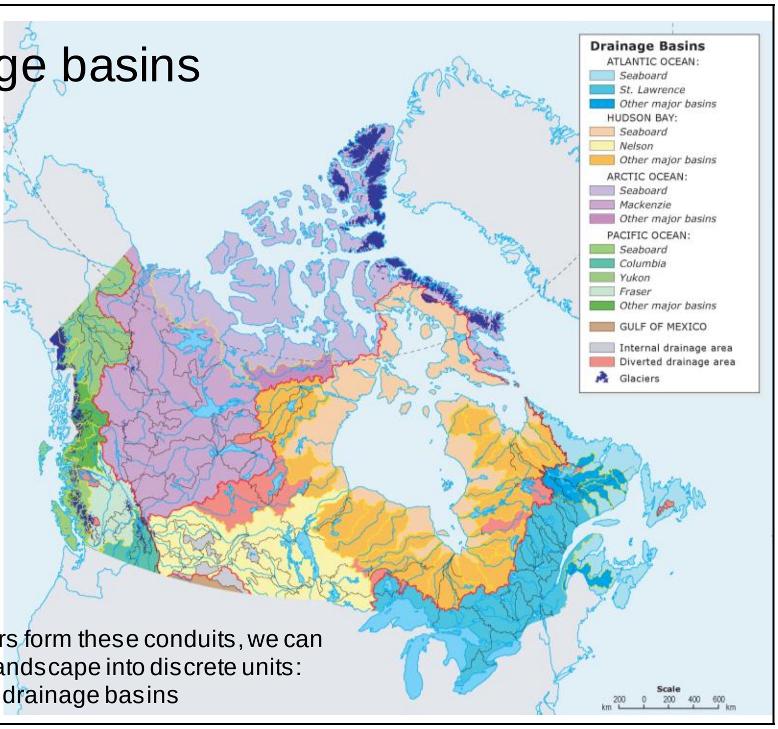
Drainage Basins

Rivers are the **conduits for water and sediment movement** in the landscape between sources and sinks.

Rivers are also conduits for the food web as well as nutrients and contaminant movement in the landscape. Thus understanding how water and sediment move through rivers is of grave importance to understanding **ecology, population dynamics, and environmental chemistry**.

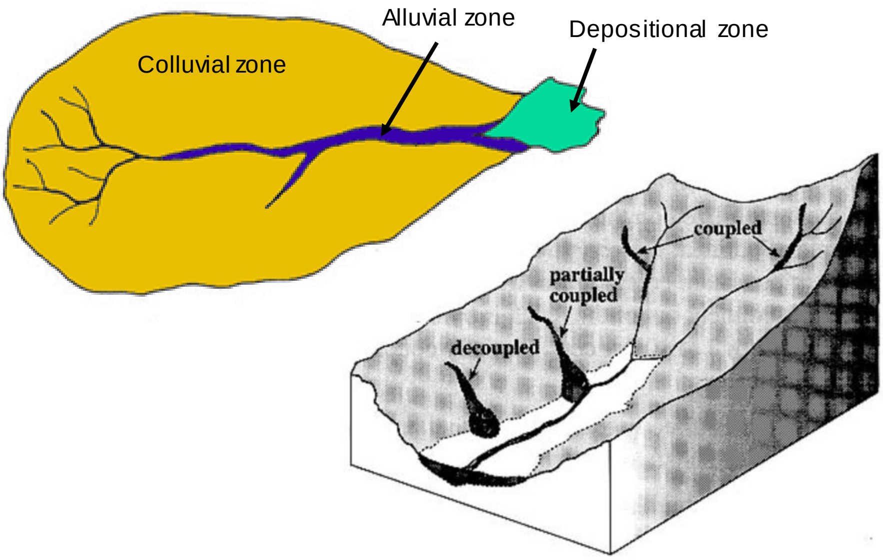
This idea is not currently well understood or accepted, simply because ecologists, biologists, and chemists are used to **thinking at small-scale**. Yet, understanding that various ‘sites’ are part of a functioning landscape and a water/sediment superhighway is increasingly recognized as **Watershed Science**.

Drainage basins



Because rivers form these conduits, we can break the landscape into discrete units: drainage basins

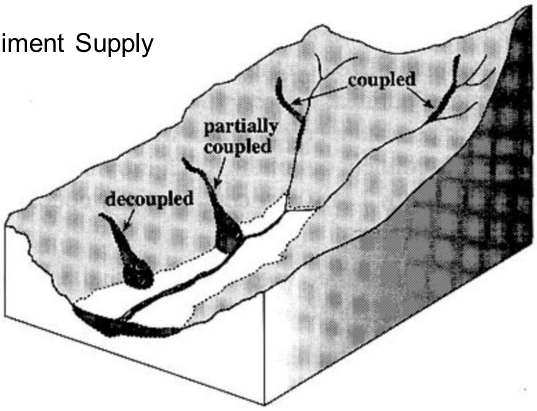
Structure of a drainage basin



Alluvial vs. Colluvial Channels

Alluvial Channels

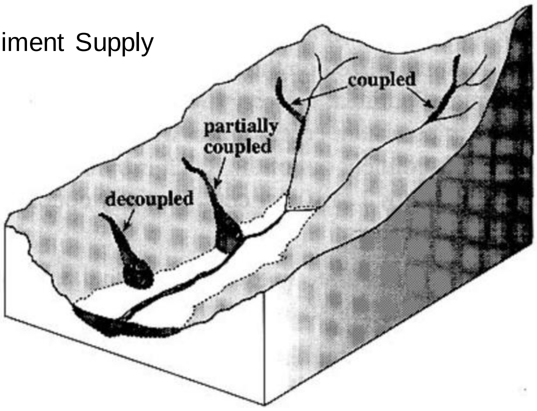
- Rivers flowing through their own deposits
- Typically lowland river channels flowing in wide valleys
- Transport Capacity ≤ Sediment Supply
- Input ≥ Output



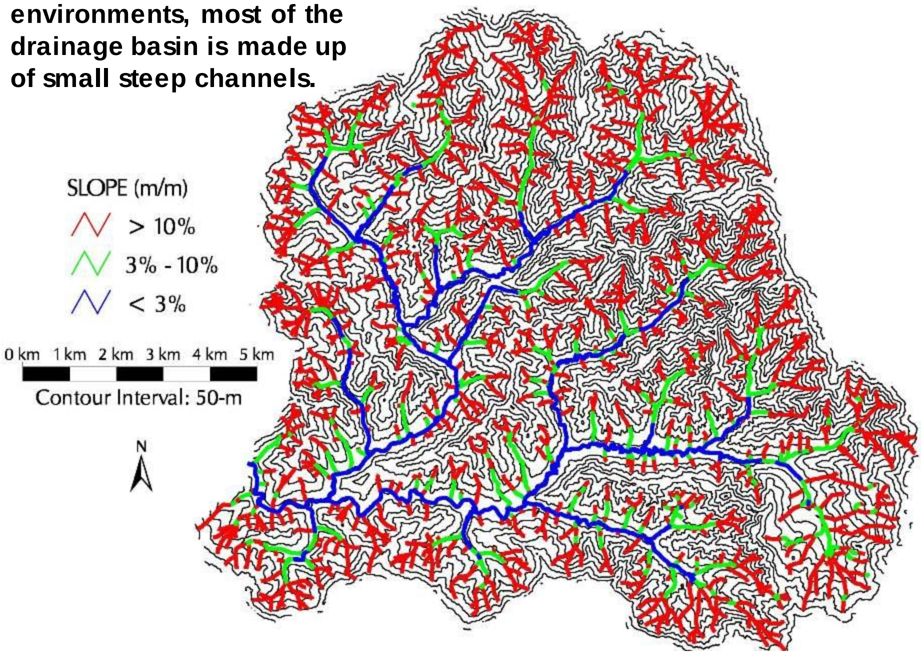
Alluvial vs. Colluvial Channels

Colluvial Channels

- Rivers actively cutting into the landscape
- Headwater channels that drain mountainous environments
- Transport Capacity > Sediment Supply



In mountainous environments, most of the drainage basin is made up of small steep channels.

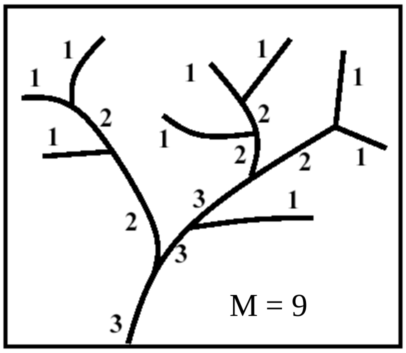


Organization of Drainage Basins

There was a time when geomorphologists thought about river systems in terms of Magnitude and Stream Order

Magnitude of a basin is the number of first order or “exterior links” in a catchment. Magnitude correctly emphasizes where the channel begins. Stream order is commonly done on nearly arbitrary network scales, and therefore means little. “Horton’s laws”, which are derived from analysis of stream orders, have no physical meaning.

Strahler stream order



N= 2M-1

N= number of links (exterior + interior)

Kirchner (Geology; 1993) correctly pointed out that stream order is an outcome of the ordering process and is statistically meaningless.

The enduring quantitative measures of drainage basins: Hack’s Law

$$D = \sum L / A$$

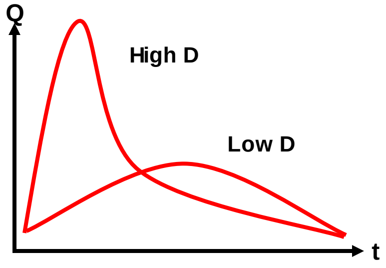
Hack’s Law: $D = \sum L / A$

Drainage density varies depending on local climate (frequency and magnitude of precipitation), vegetation patterns, and geology.

Importance arises because:

1) It is a real measure of drainage basin structure, not dependent on ordering schemes.

2) The time scales related to rainfall concentration and runoff are related to D.



3) Can be used to define other useful geomorphic properties of the landscape: ***constant of channel maintenance*** and ***length of overland flow***

Length-area relation
$L = 1.4A^{0.6}$

Hack’s Law: $D = \sum L / A$

constant of channel maintenance:

$$C = 1 / D = A / \sum L$$

Basin area required to maintain one linear unit of channel

length of overland flow:

$$LOF = 1 / 2D$$

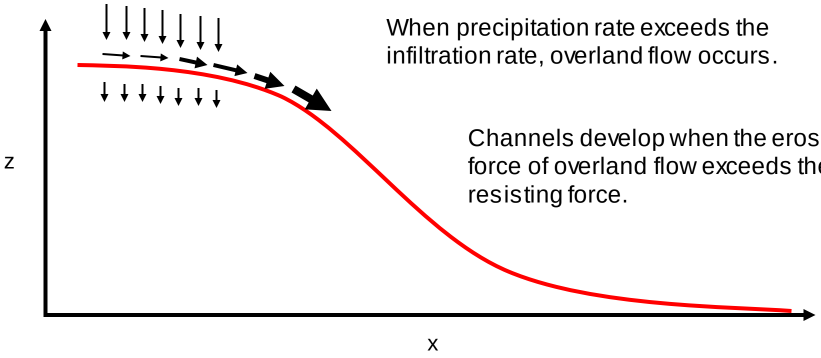
Distance of water movement on a hillslope before it begins to channelize

	S. California	Utah	Indiana	Virginia
	Igneous/ Meta.	Sedimentary	Sandstone	Metamorphic
D (km/km^2)	13.70	5.58	3.83	2.3
C (m^2/m)	73	179	261	435
LOF (m)	36	90	131	217

What controls drainage density?

Channel head: threshold transition between hillslope and channel processes

Hortonian model for channel initiation

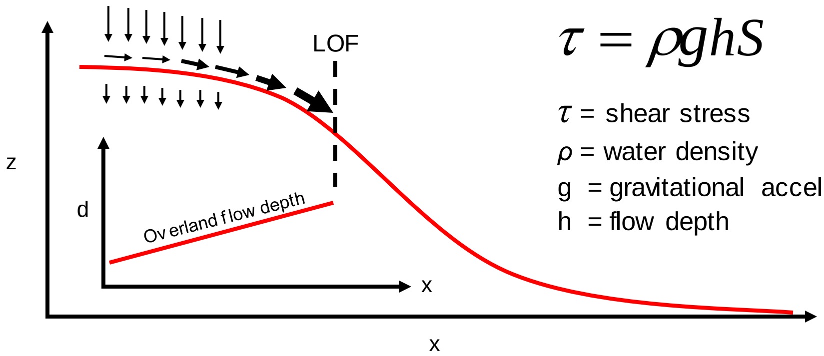


When precipitation rate exceeds the infiltration rate, overland flow occurs.

Channels develop when the erosive force of overland flow exceeds the resisting force.

The ***resisting force*** is given by the weight of a grains resting on the surface of the slope + and cohesive forces (clay, veg, roots, etc.)

Hortonian model for channel initiation



$$\tau = \rho g h S$$

τ = shear stress
ρ = water density
g = gravitational accel.
h = flow depth

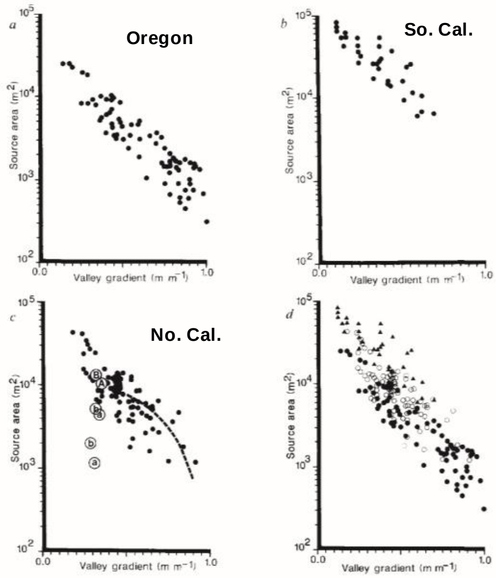
Moving downslope, the volume of flowing water increases (and so does the slope), so the ***erosive force*** also increases until it can move sediment.

Montgomery and Dietrich (Nature, 1988)

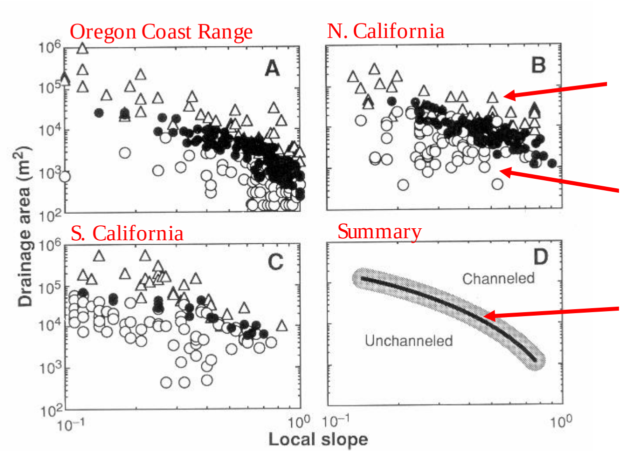
Explored the problem of where channels begin by examining the relation between the slope and upslope source area for water and sediment at channel heads.

They found an inverse slope between contributing area and slope – as slope increases, the drainage area declines (as in Hortonian model)

But, they found that most channels are initiated at sites of landslides in steep terrain suggesting channels are initiated by hillslope processes not channel processes.



Montgomery and Dietrich (Science, 1992)



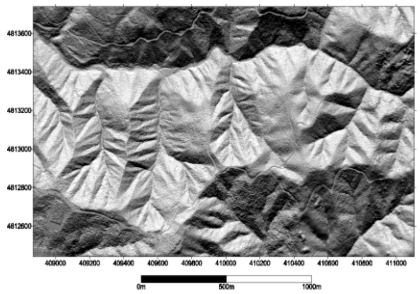
Downslope of channel head

Upslope of channel head

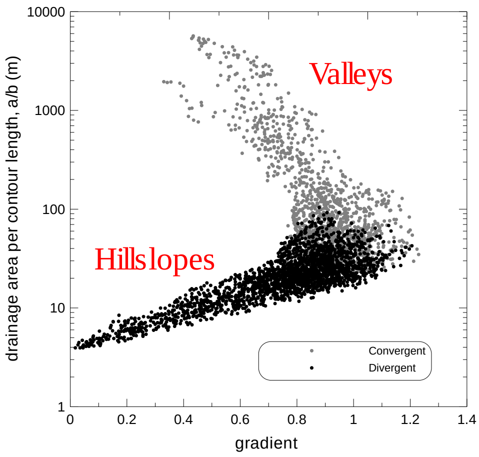
Threshold for channel initiation

They also found that for a given slope, there is a threshold contributing area for channel initiation. Below the threshold, hillslope processes dominate. Above the threshold, channel processes dominate.

Diffusive and advective landscape elements

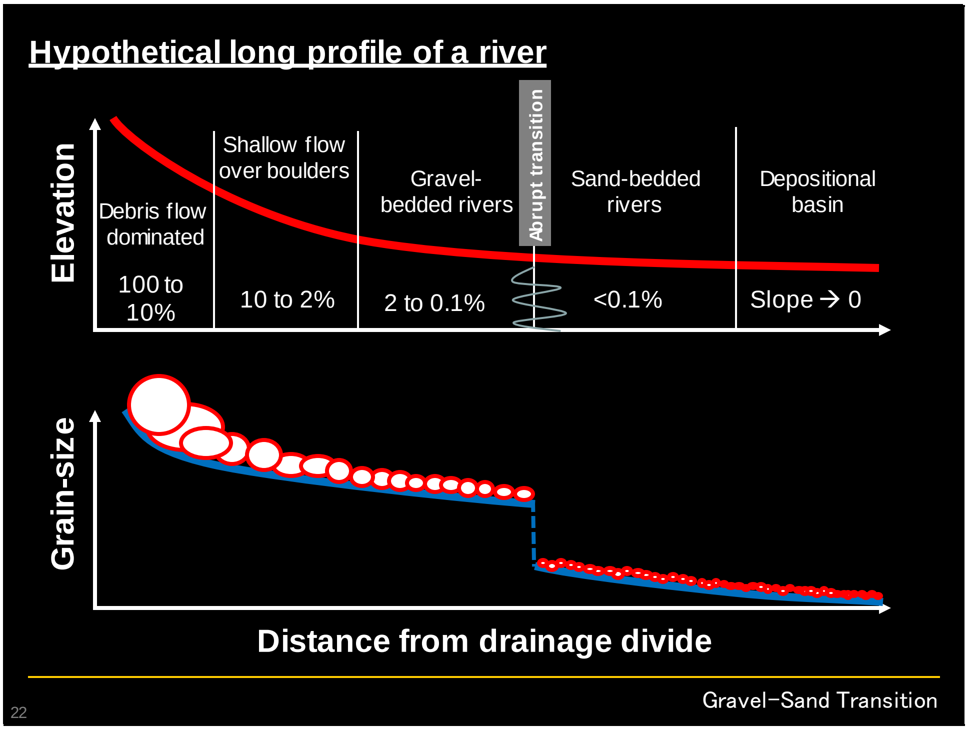


Starting at the crest of a hill, the drainage area increases with slope. At the hillslope-valley transition, the slope begins to declines and the drainage area increases (big rivers have small gradients).



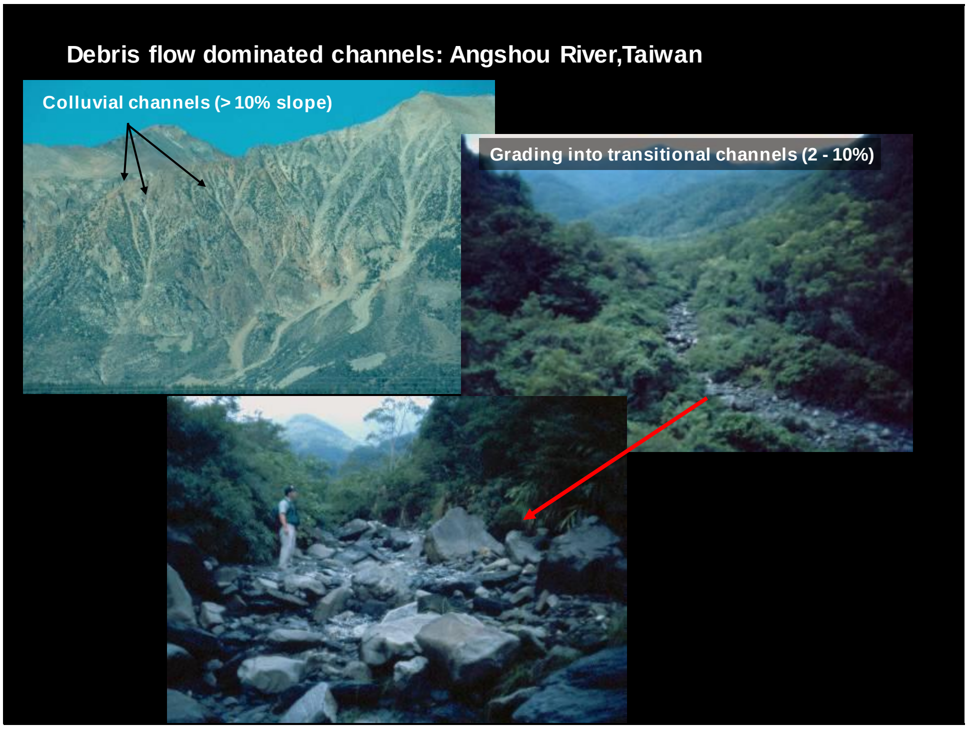
Dietrich et al. (PIG, 2003)

Long profile and downstream fining



Hypothetical long profile of a river

Gravel–Sand Transition



Debris flow dominated channels: Angshou River,Taiwan

Colluvial channels (> 10% slope)

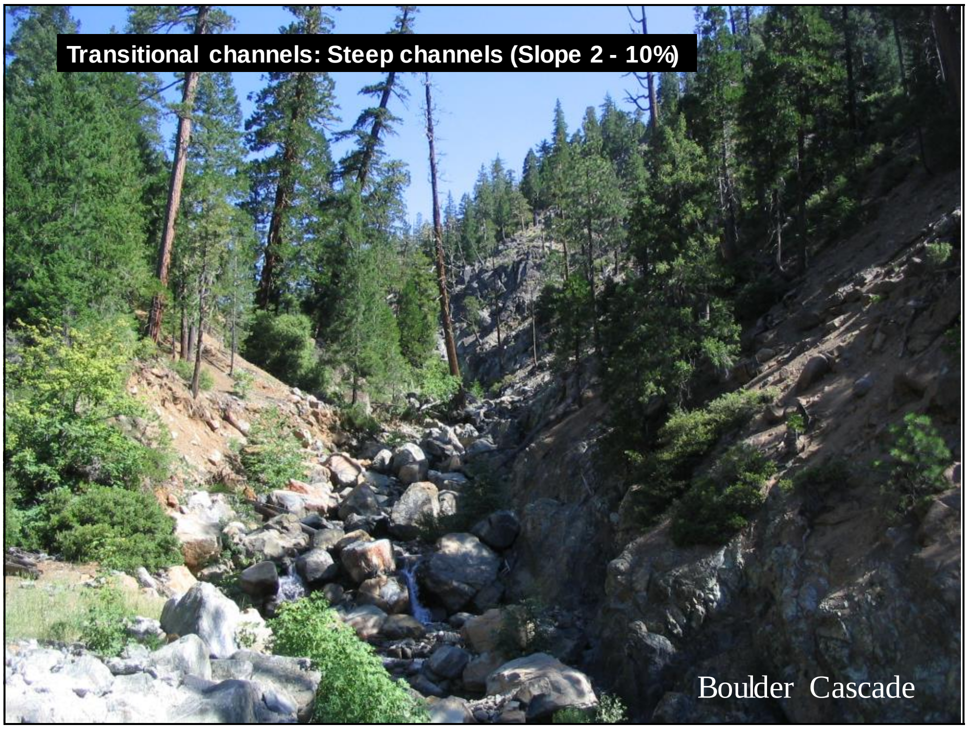
Grading into transitional channels (2 - 10%)

Transitional channels: Steep channels (Slope 2 - 10%)

Erlenbach Torrent
Switzerland: 10% slope

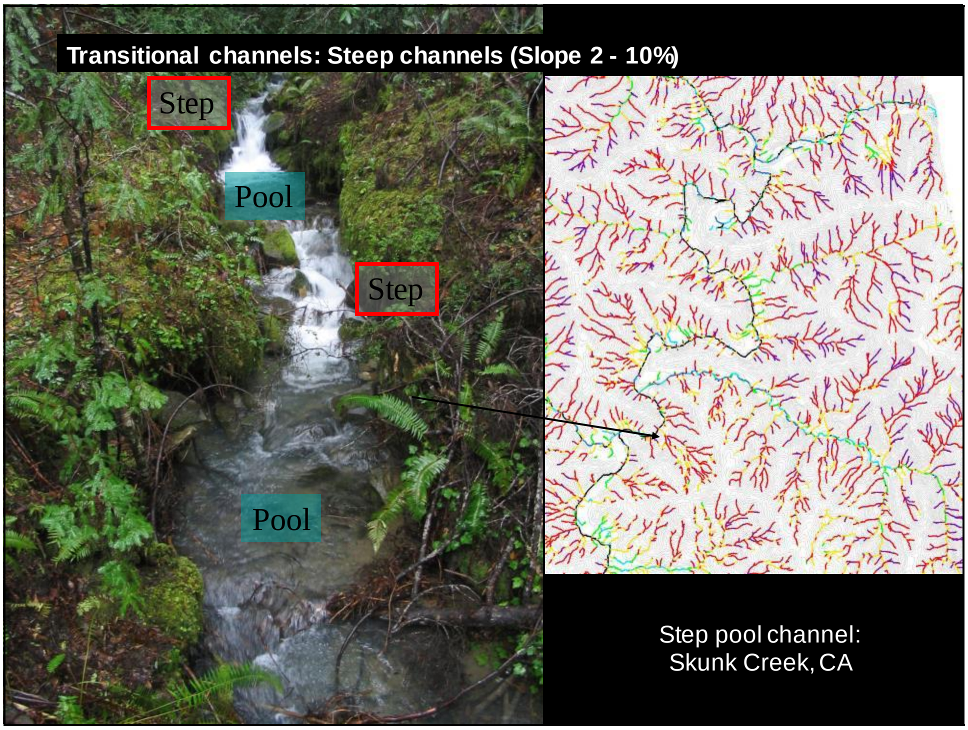
Finer, more mobile sediment

Large, relatively immobile grains



Transitional channels: Steep channels (Slope 2 - 10%)

Boulder Cascade



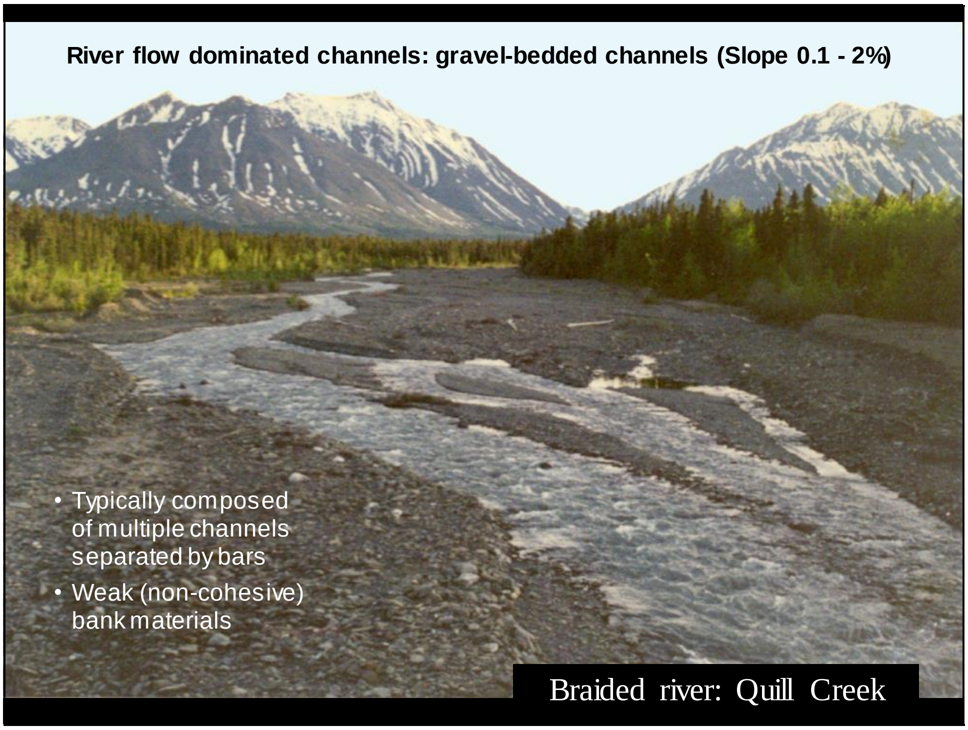
Transitional channels: Steep channels (Slope 2 - 10%)

Step pool channel:
Skunk Creek, CA

Transitional channels: Steep channels (Slope 2 - 10%)

River flow dominated channels: gravel-bedded channels (Slope 0.1 - 2%)

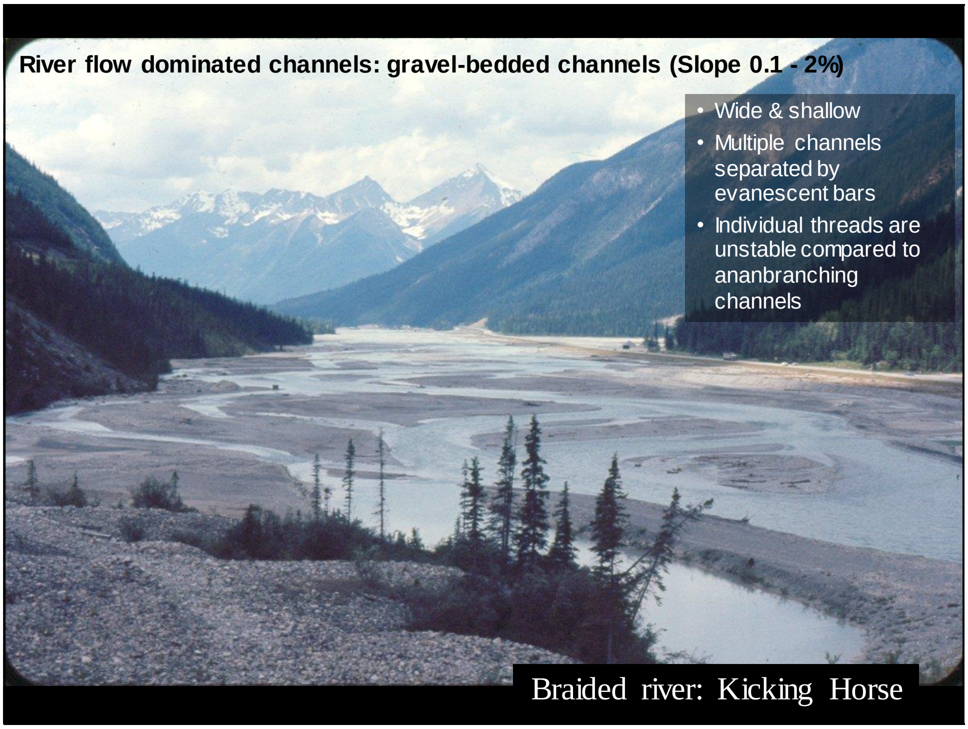
River flow dominated channels: gravel-bedded channels (Slope 0.1 - 2%)



River flow dominated channels: gravel-bedded channels (Slope 0.1 - 2%)

- Typically composed of multiple channels separated by bars
- Weak (non-cohesive) bank materials

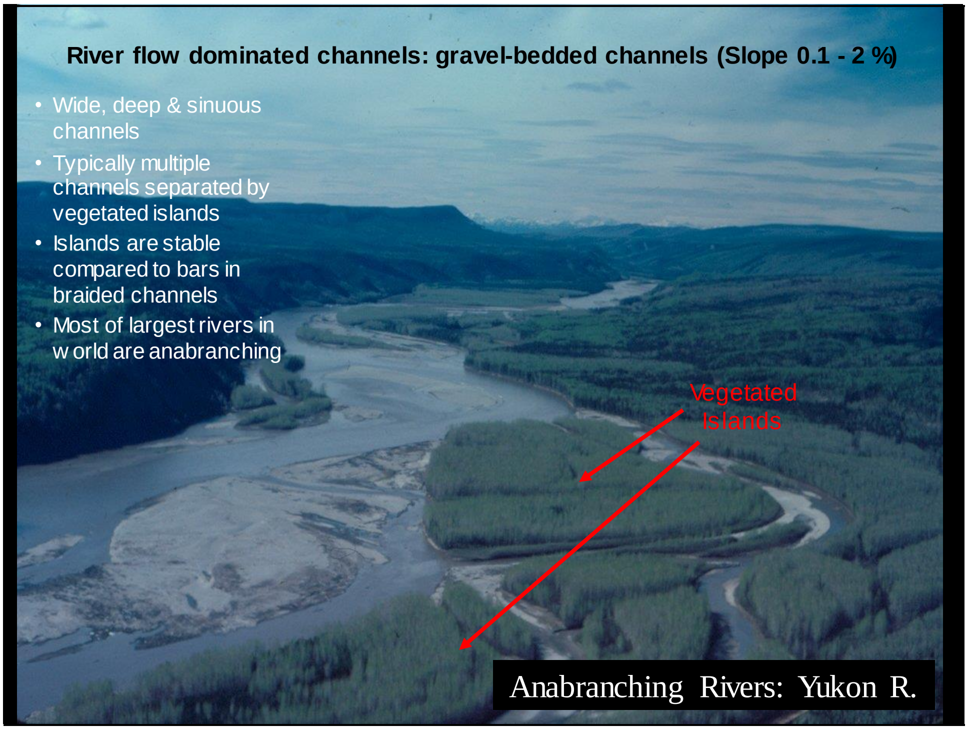
Braided river: Quill Creek



River flow dominated channels: gravel-bedded channels (Slope 0.1 - 2%)

- Wide & shallow
- Multiple channels separated by evanescent bars
- Individual threads are unstable compared to ananbranching channels

Braided river: Kicking Horse

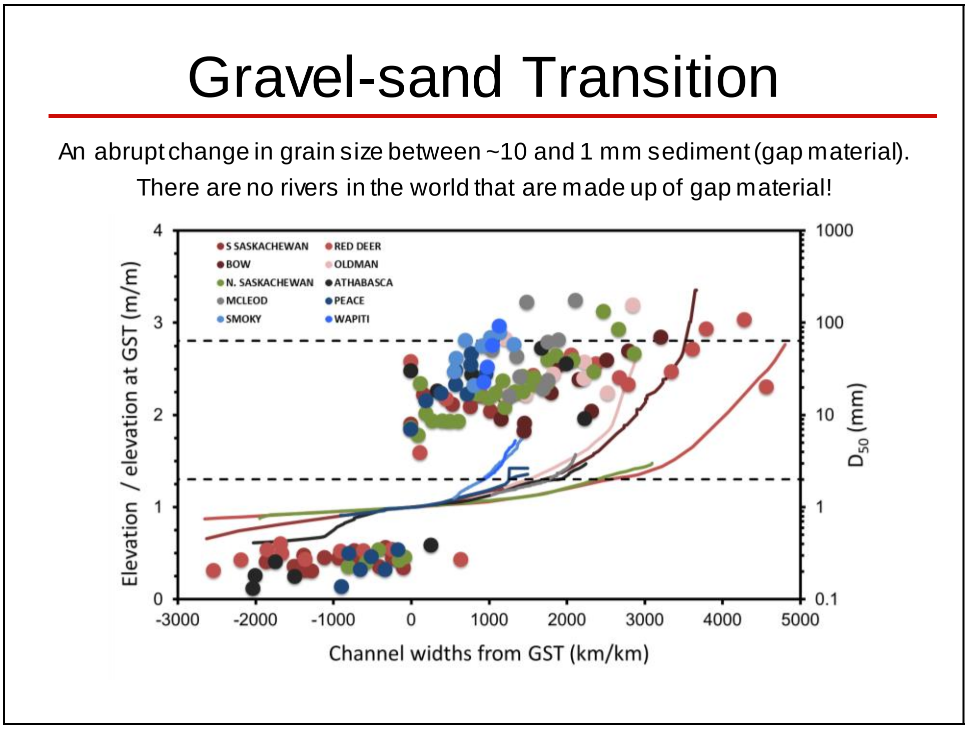


River flow dominated channels: gravel-bedded channels (Slope 0.1 - 2 %)

- Wide, deep & sinuous channels
- Typically multiple channels separated by vegetated islands
- Islands are stable compared to bars in braided channels
- Most of largest rivers in world are anabranching

Vegetated Islands

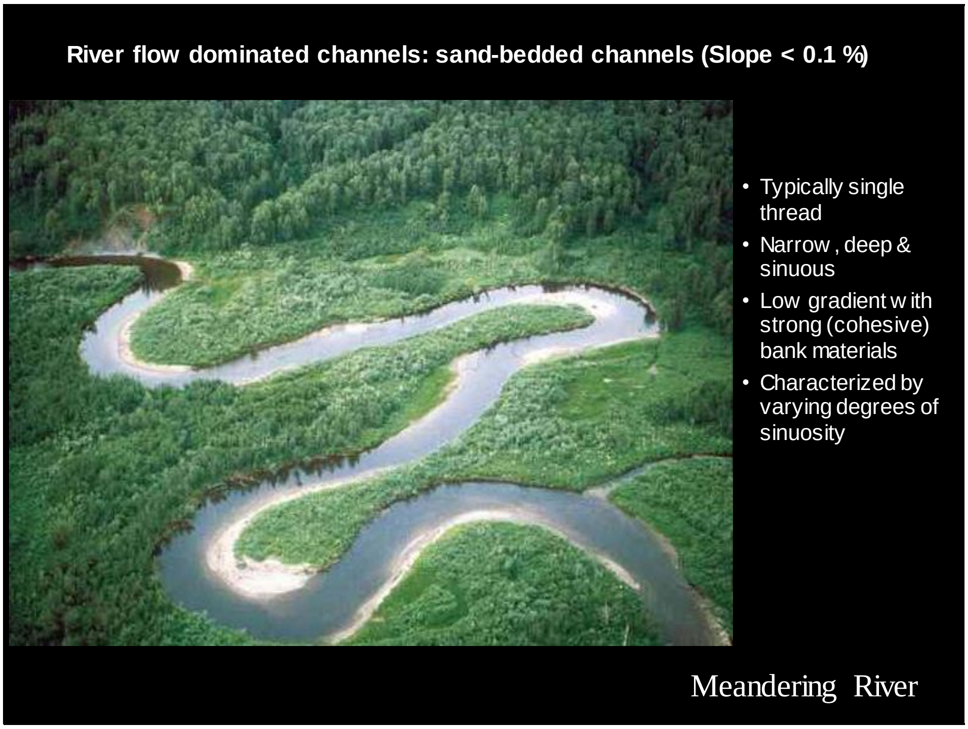
Anabranching Rivers: Yukon R.



Gravel-sand Transition

An abrupt change in grain size between ~10 and 1 mm sediment (gap material). There are no rivers in the world that are made up of gap material!

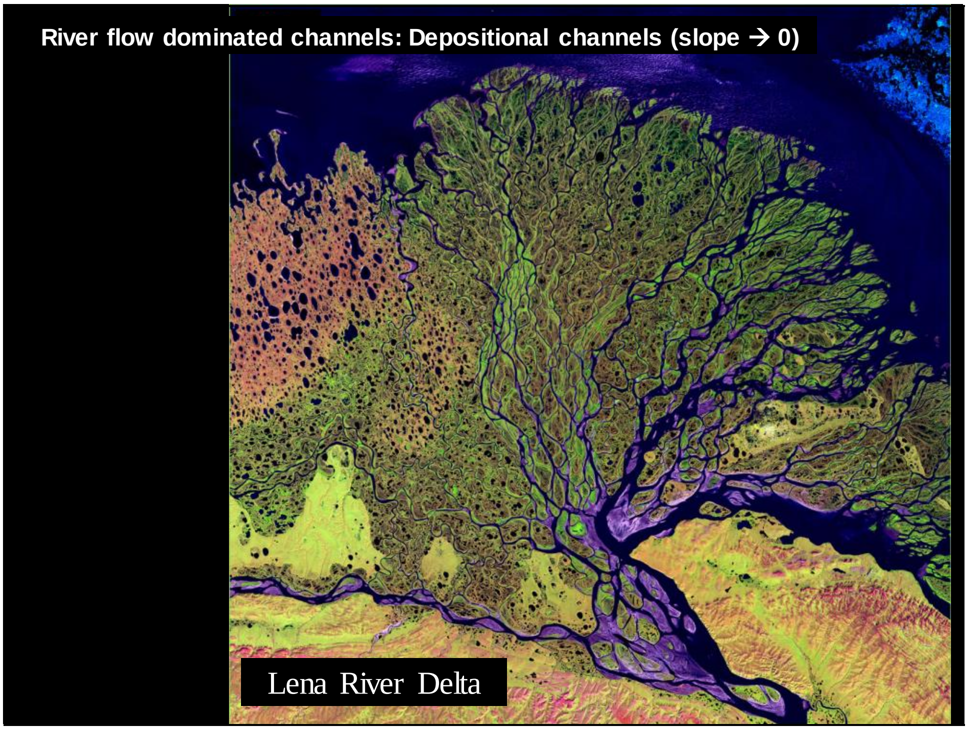
River flow dominated channels: sand-bedded channels (Slope < 0.1 %)



River flow dominated channels: sand-bedded channels (Slope < 0.1 %)

- Typically single thread
- Narrow, deep & sinuous
- Low gradient with strong (cohesive) bank materials
- Characterized by varying degrees of sinuosity

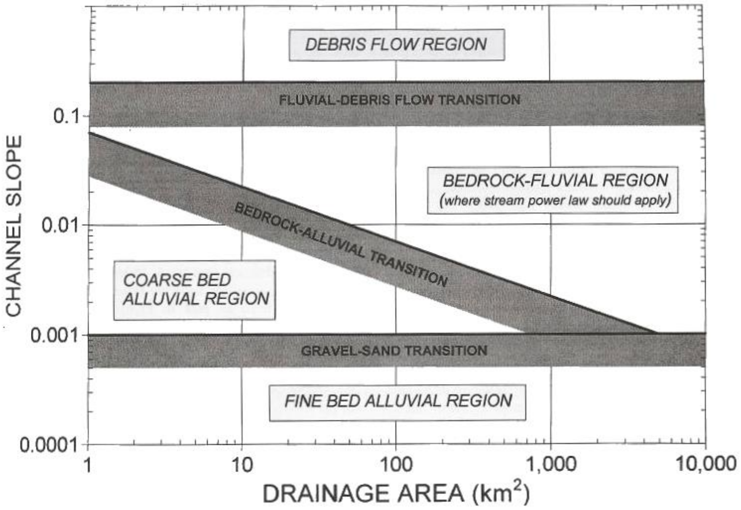
Meandering River



River flow dominated channels: Depositional channels (slope → 0)

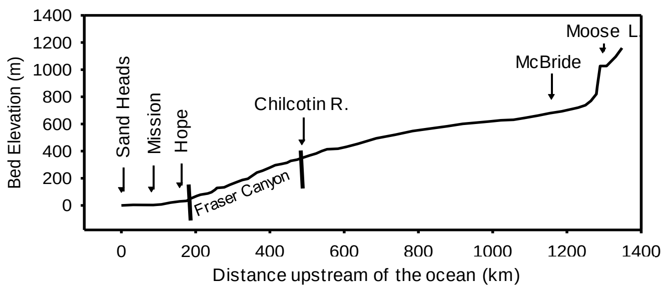
Lena River Delta

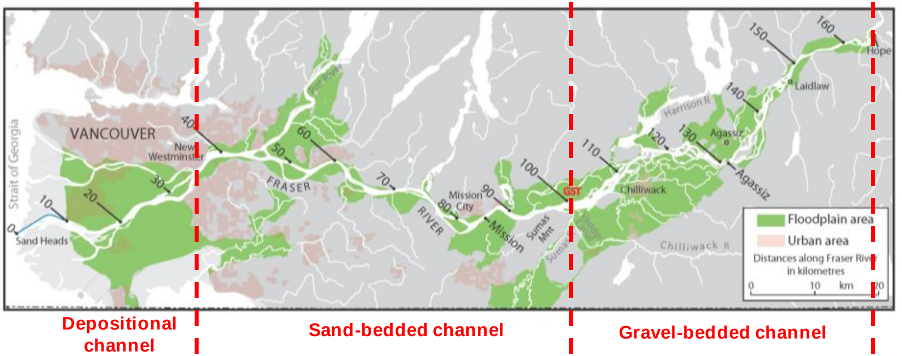
Downstream grainsize change

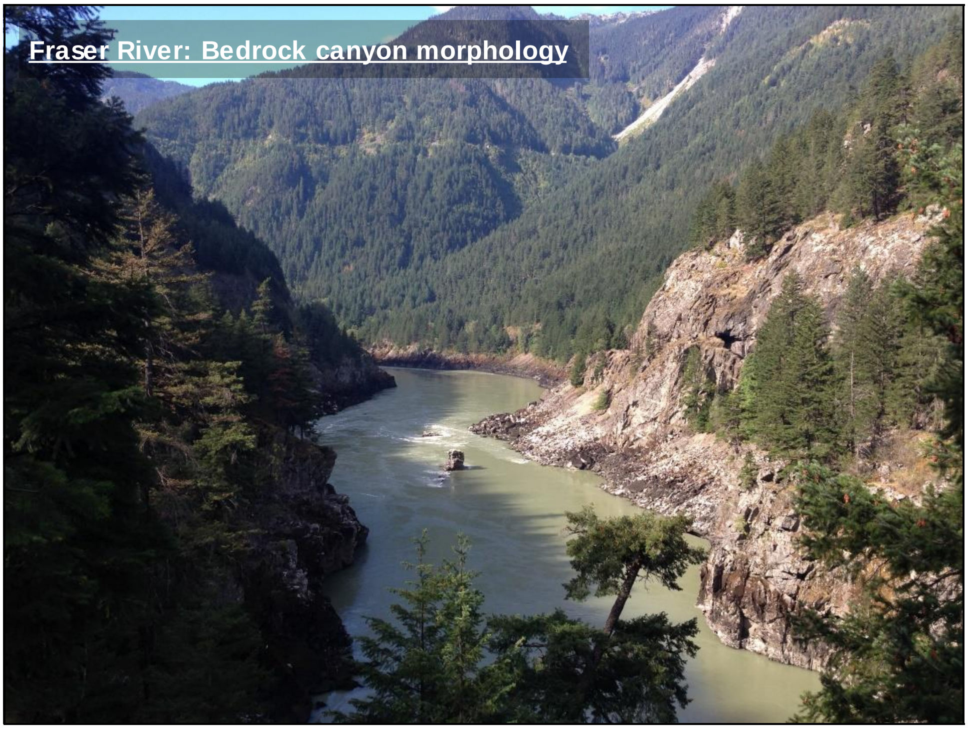


Sklar and Dietrich (2001)

Fraser River, BC







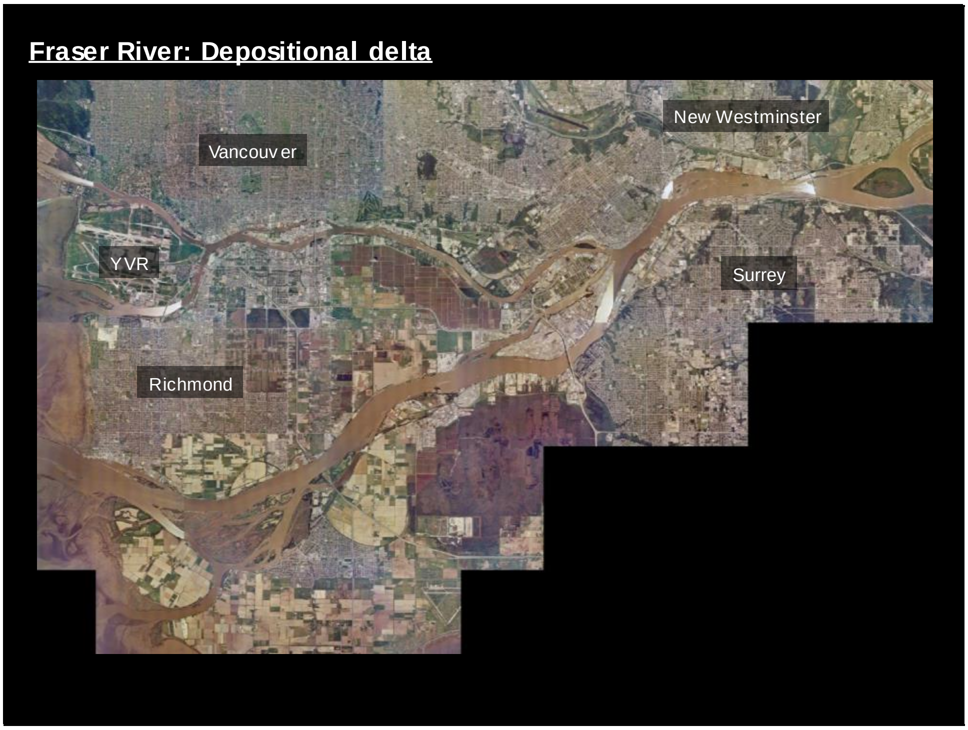
Fraser River: Bedrock canyon morphology

Fraser River: Multi-thread gravel-bedded anabranching morphology

Andre Zimmermann

Fraser River: Single-thread sand-bedded ‘pseudo-meandering’

Most of the water is not carried through the apex of the bends!

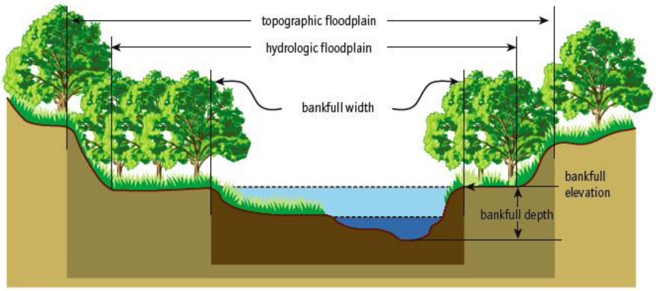


Fraser River: Depositional delta

River channel morphology

Floodplains & bankfull flow

The floodplain is that *part of the river* that is inundated whenever flow exceeds the bank height.

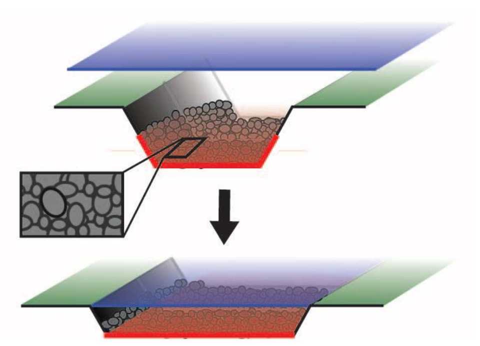


The flow that is just contained within the river channel banks without flowing on to the floodplain is the bankfull flow.

The return interval for bankfull flow is ever couple of years (1.5 – 2.5).

Bankfull flow is the channel forming flow.

Formative and bankfull flows



Phillips and Jerolmack, Science 2016

- Bankfull flow sets the channel dimensions in alluvial rivers.
- Flows much above the threshold for motion of the bed cause channels to readjust their size
- In gravel-bed rivers, bankfull flows just exceed the threshold of motion
- The morphology of alluvial channels is self-regulated
- Alluvial channels are ‘authors of their own geometry’ – Gary Parker.

Hydraulic Geometry of Rivers

Width $w = aQ^b$

Depth $d = cQ^f$

Velocity $U = kQ^m$

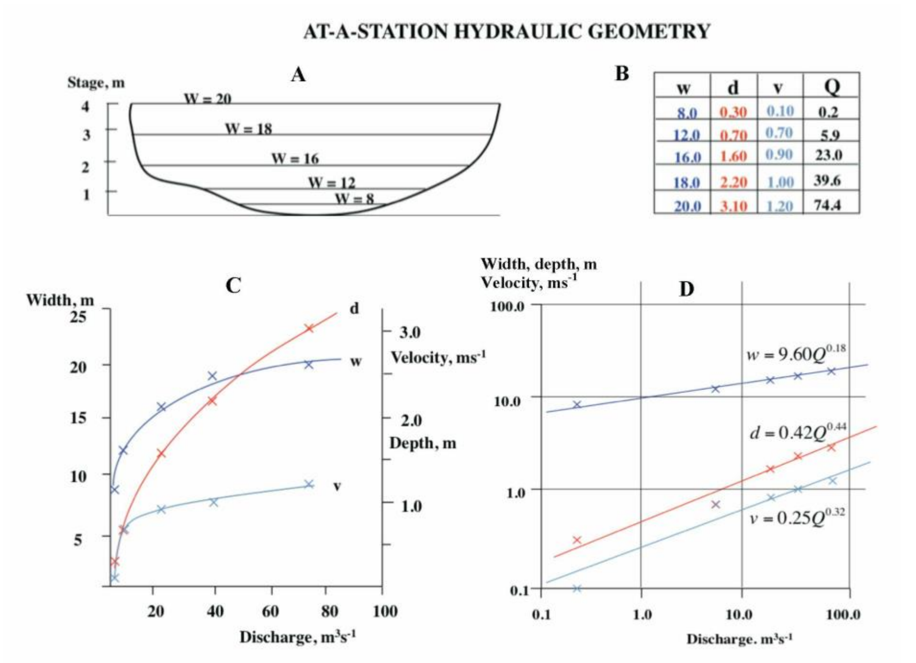
***At-a-station:* tells us the channel dimensions as flow changes at one cross-section**

***Downstream:* tells us how channel dimensions change along the channel.**

$$b + f + m = 1$$
$$a * c * k = 1$$

Luna Leopold:

served as Chief of the Hydrology Section of the USGS in the late 1950s and 1960s. There he had access to an enormous data set that gave the dimensions (width, depth) and river channel velocity at various discharges and along stream. Invented what we now call Hydraulic Geometry.



w	d	v	Q
8.0	0.30	0.10	0.2
12.0	0.70	0.70	5.9
16.0	1.60	0.90	23.0
18.0	2.20	1.00	39.6
20.0	3.10	1.20	74.4

Ted Hickin course notes

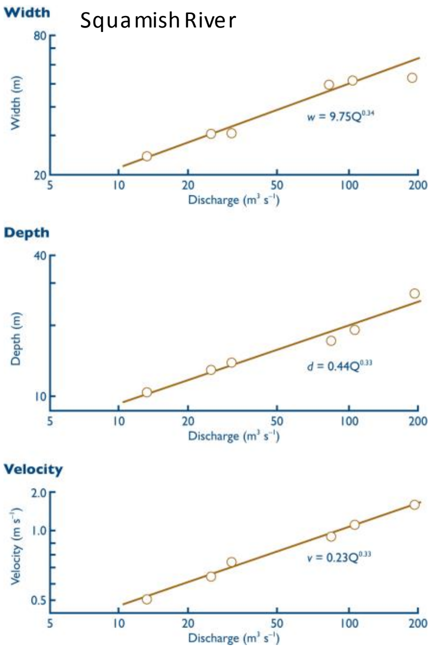
At-A-Station Hydraulic Geometry of Rivers

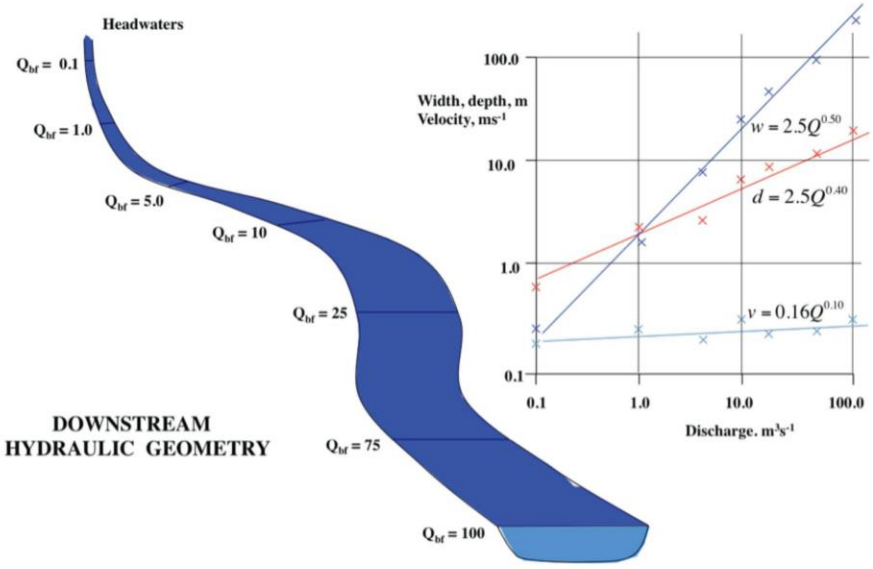
Width $w = aQ^b$

Depth $d = cQ^f$

Velocity $U = kQ^m$

	At-a-station		
Channel locality and type	**b**	**F**	**m**
Midwest USA	0.26	0.40	0.34
Midwest USA			
Ephemeral streams, semi-arid USA	0.25	0.41	0.33
Upper Salmon River (Idaho)			
R. Bollin Dean, coarse-bed cohesive banks	0.12	0.40	0.48
British gravel bed rivers			
Columbia River (BC), canal-like anastomosed sand channels, cohesive banks	0.10	0.66	0.24


Squamish River



Ted Hickin course notes

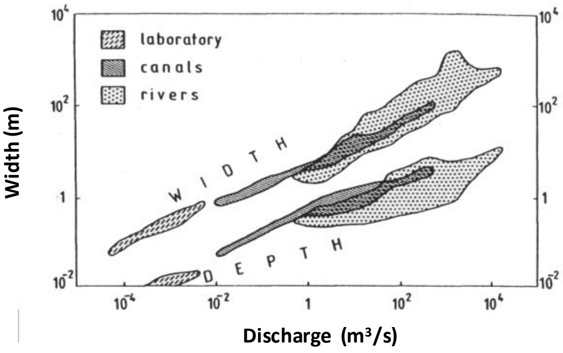
Downstream Hydraulic Geometry of Rivers

Width $w = aQ^b$

Depth $d = cQ^f$

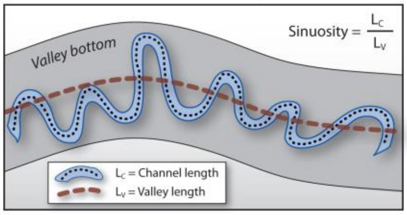
Velocity $U = kQ^m$

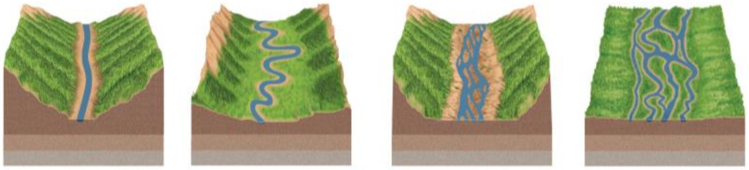
	Downstream values		
Channel locality and type	b	F	m
Midwest USA	0.50	0.40	0.10
Midwest USA	0.46	0.38	0.16
Ephemeral streams, semi-arid USA	0.50	0.30	0.20
Upper Salmon River (Idaho)	0.54	0.34	0.12
R. Bollin Dean, coarse-bed cohesive banks	0.46	0.16	0.38
British gravel bed rivers	0.45	0.40	0.15
Columbia River (BC), canal-like anastomosed sand channels, cohesive banks			



Alluvial river channel planforms

Channel sinuosity





Straight
sinuosity < 1.3

Meandering
sinuosity > 1.5

Braided
unvegetated bars

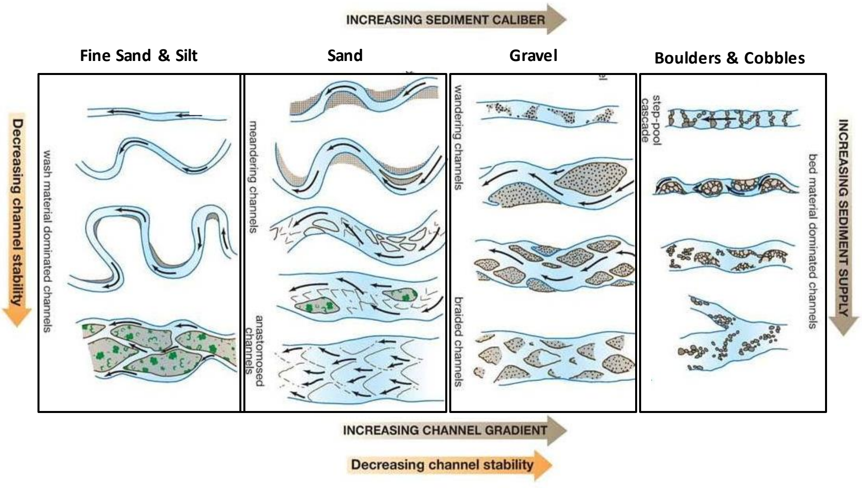
Anastomosing
vegetated islands

Single-thread channels exhibit either **straight** or **meandering** patterns. Straight channels are typically confined by valley walls and have **sinuosity** < 1.3, whereas meandering channels typically flow across broad floodplains and have sinuosity > 1.5. The meander belt is the zone of active meandering.

Multi-thread channels exhibit either **braided** or **anastomosing** patterns. Braided channels exhibit multiple unvegetated, frequently shifting channels that converge and diverge within a larger channelway. Anastomosing channels divide into multiple channels that flow around vegetated islands.

Bierman & Montgomery text

What controls river planform?

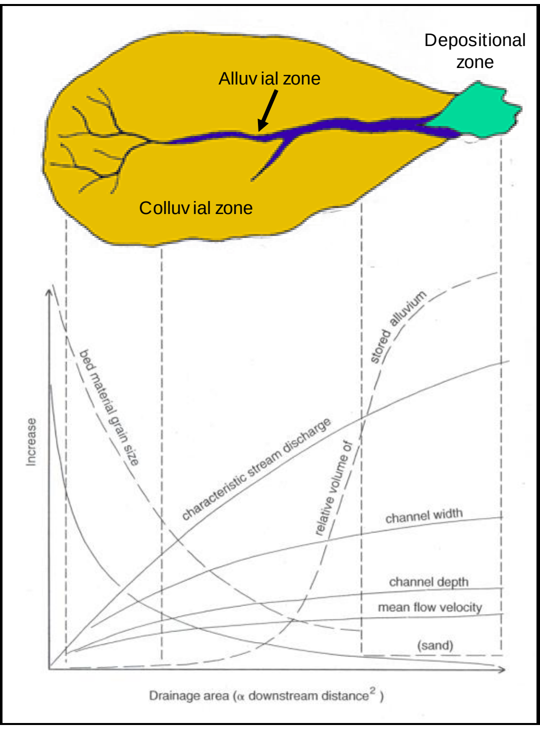


Interplay between sediment size, sediment supply and channel gradient

Church, Annual Reviews of Earth Science, 2005

Summary of downstream changes in rivers

- Elevation and slope decline
- Rivers transition from bedrock to alluvial
- The amount of stored alluvium increases
- Bed material grain size declines with an abrupt transition between gravel and sand
- Rivers get bigger (width, depth, velocity)



Goals of Rivers Lectures

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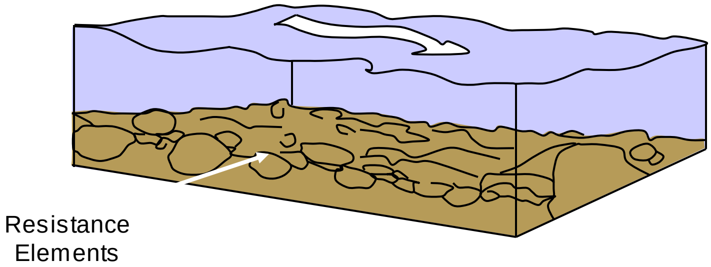
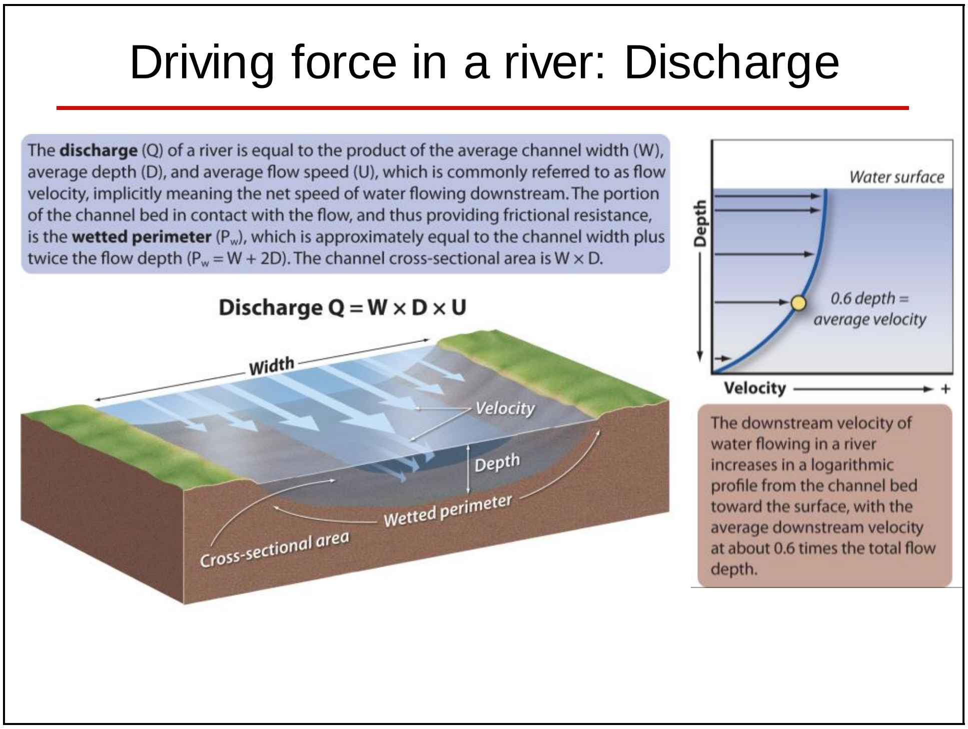
Flow in rivers

River Processes

- River channels are stable features in the landscape because of a balance between the **downstream driving force (gravity)** which causes flow and **resistance to flow**
- This resistance to down-slope flow is communicated to the fluid via the generation of turbulent flow eddies at the boundary



- Turbulent eddies extract energy from the mean flow and convert it to turbulent energy maintaining the balance



Driving force in a river: Discharge

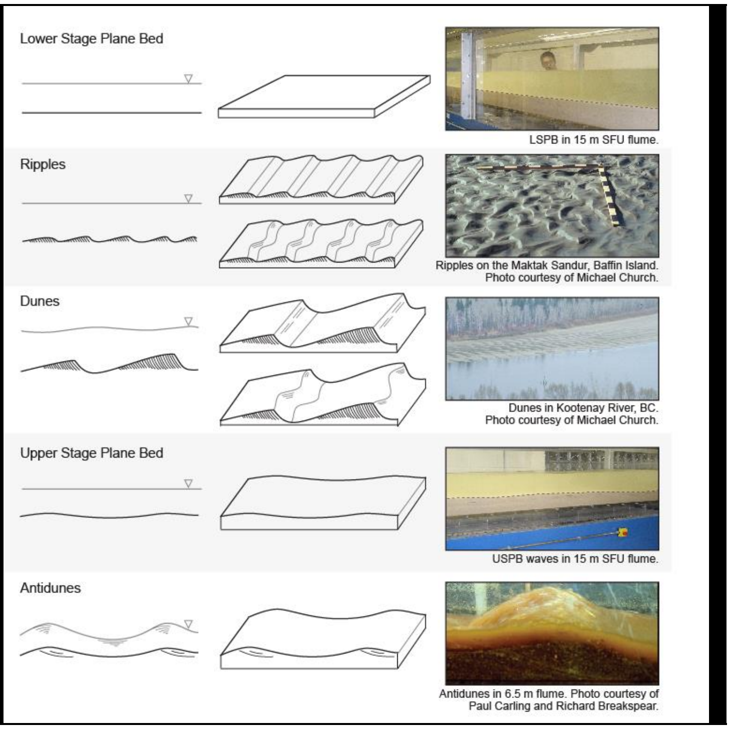
The **discharge** (Q) of a river is equal to the product of the average channel width (W), average depth (D), and average flow speed (U), which is commonly referred to as flow velocity, implicitly meaning the net speed of water flowing downstream. The portion of the channel bed in contact with the flow, and thus providing frictional resistance, is the **wetted perimeter** (P_w), which is approximately equal to the channel width plus twice the flow depth ($P_w = W + 2D$). The channel cross-sectional area is W × D.

Discharge $Q = W \times D \times U$

The downstream velocity of water flowing in a river increases in a logarithmic profile from the channel bed toward the surface, with the average downstream velocity at about 0.6 times the total flow depth.

Resisting force: frictional drag

Flow Resistance in Sand-Bedded Rivers



Flow Resistance Equations

In practice, flow resistance is often calculated in terms of its influence on velocity.

Engineers and scientists have attempted to devise methods for calculating velocity for centuries. This has resulted in a number of expressions for the mean velocity in rivers:

Manning **Chezy** **Darcy-Wiesbach**

$$U = \frac{d^{2/3} S^{1/2}}{n} = C d^{1/2} S^{1/2} = 8\left(\frac{gdS}{f}\right)^{1/2}$$

Each equation includes the flow depth (d), the river slope (S) and a flow resistance coefficient (*n, C, f*).

Flow Resistance Equations

A significant difference between these equations is that in the **Manning Equation**:

$$U \propto d^{2/3} \mathrm{S}^{1/2}$$

While in the **Chezy and D-W** equations:

$$U \propto d^{1/2} \mathrm{S}^{1/2}$$

The Manning equation is empirical (based on observations), so it is widely thought to have proportionality correct.

Flow Resistance Equations

There is a significant problem with the Manning and Chezy equations in that the coefficients have units (n = [$L^{1/6}$] and C = [$^{1/2}$]). This means that roughness is dependent on the size of the river, which we know is not correct.

The Darcy-Weisbach Equation is increasingly used to replace the Manning and Chezy. You can also modify the Manning or Chezy equations so that C or n is dependent on the river bed grain-size:

$$C = 8\left(\frac{d}{k_s}\right)^{1/6}\sqrt{g} \qquad n = \frac{1}{8\,k_s^{-1/6}\sqrt{g}}$$

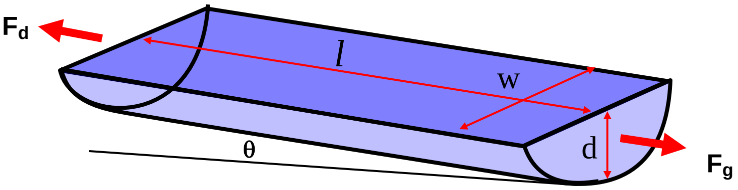
Where $k_s = 2\ D_{90}$ in gravel-bedded channels, $2\ D_{50}$ in flat sand-bedded rivers, and is proportional to the bedform size in rough sand-bedded rivers.

Using the Manning or Chezy equation without an indexed coefficient should not be done!

Balance between driving and resisting forces

The balance between the resisting forces and driving forces in a river reach can be expressed mathematically by stating:

Driving force (F_g) = Resisting force (F_d)



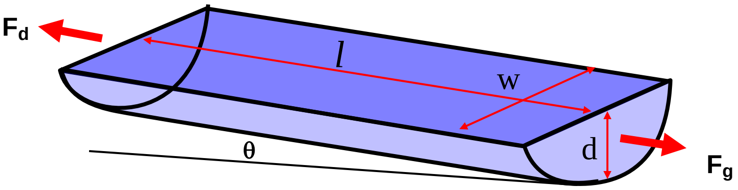
Driving force

$$F_g = mg\sin\theta = \rho(d\,l\,w)\,g\sin\theta$$

Balance between driving and resisting forces

The resisting force is calculated as the shear stress exerted on the channel boundary:

The channel boundary is the part of a river reach that is in contact with water



Resisting force: $F_d = \tau[l\,(2d + w)]$

Quantifying the Driving Force in Rivers

By equating the driving resisting force to the resisting force, we can find an expression for the driving force (boundary shear stress):

$$\rho(d\cancel{l}w)g\sin\theta = \tau[\cancel{l}(2d+w)]$$

Solving for τ

$$\tau = \rho\frac{wd}{2d+w}g\sin\theta$$

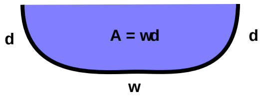
Quantifying the Driving Force in Rivers

$$\rho(d\cancel{l}w)g\sin\theta = \tau[\cancel{l}(2d+w)]$$

Solving for τ

$$\tau = \rho\frac{wd}{2d+w}g\sin\theta$$

Defining the hydraulic radius: $R = \frac{wd}{2d+w}$



Boundary shear stress in rivers → $\boxed{\tau = \rho g R\sin\theta}$

Quantifying the Driving Force in Rivers

In most river channels, when the width is greater than ~20 times the depth, the hydraulic radius is within 10% of the depth:

So: $$\tau = \rho g d \sin\theta$$

At slopes <10% or 5.7°, sin θ ≈ tan θ ≡ S

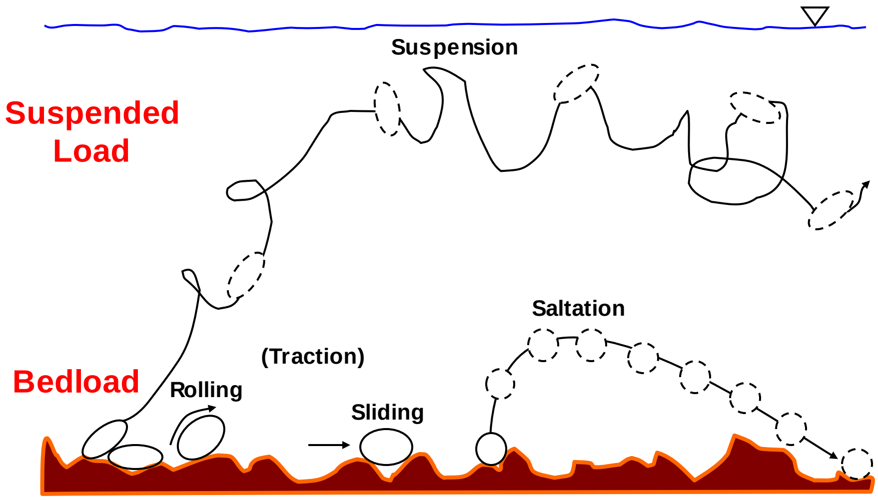
So: $$\tau = \rho g d S$$

The force exerted on the bed by the flow is equal to the flow depth times the channel slope.

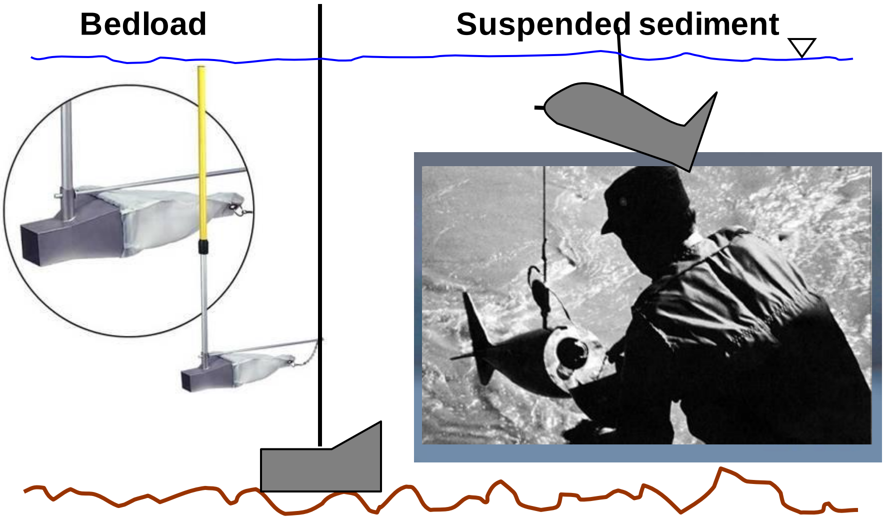
We can use the boundary shear stress to predict sediment transport in rivers.

Sediment transport in rivers

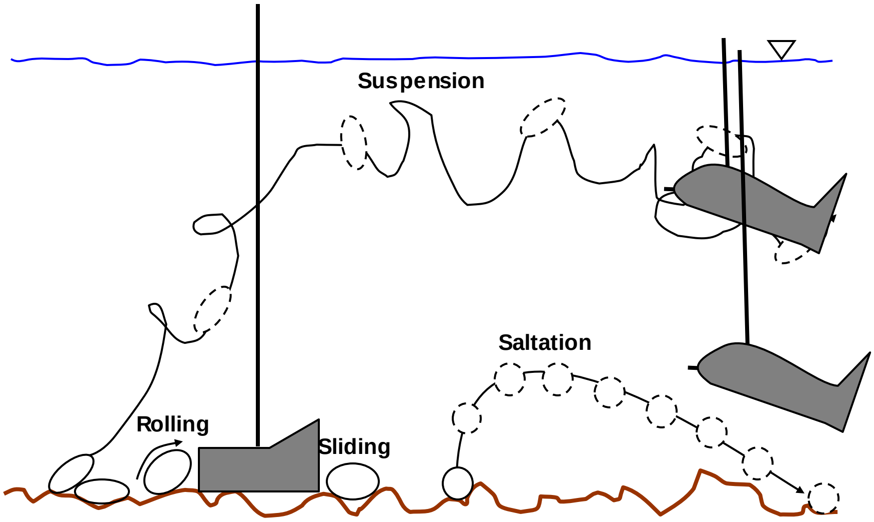
Sediment Transport Mechanisms



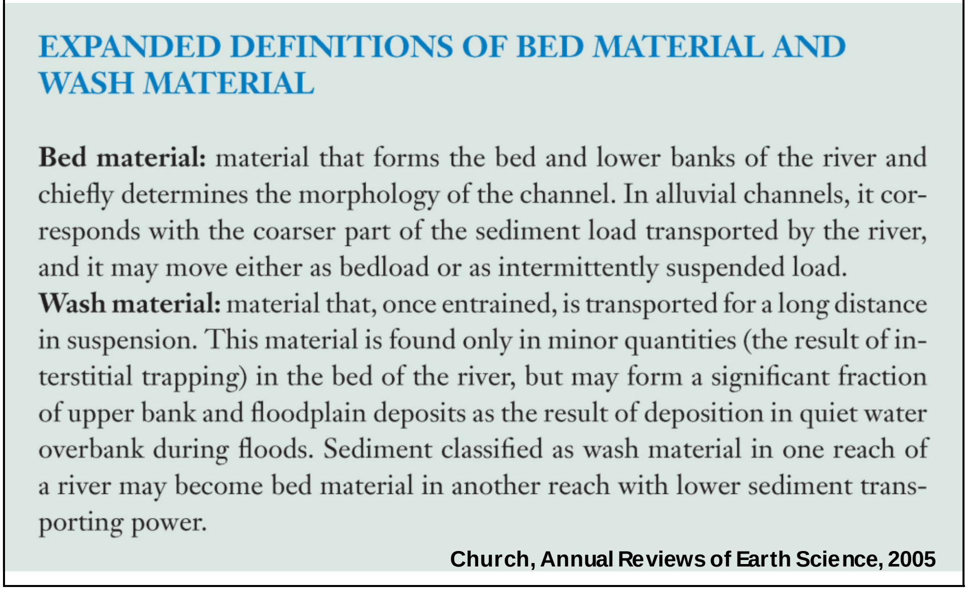
Sediment Transport Measurement in Rivers



Mechanism and Measurement Discrepancy



Sediment Transport in Rivers



EXPANDED DEFINITIONS OF BED MATERIAL AND WASH MATERIAL

Bed material: material that forms the bed and lower banks of the river and chiefly determines the morphology of the channel. In alluvial channels, it corresponds with the coarser part of the sediment load transported by the river, and it may move either as bedload or as intermittently suspended load.

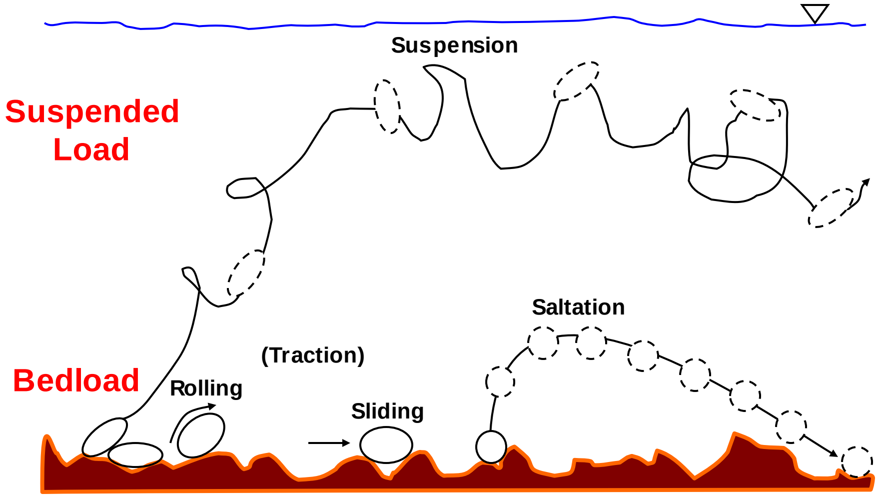
Wash material: material that, once entrained, is transported for a long distance in suspension. This material is found only in minor quantities (the result of interstitial trapping) in the bed of the river, but may form a significant fraction of upper bank and floodplain deposits as the result of deposition in quiet water overbank during floods. Sediment classified as wash material in one reach of a river may become bed material in another reach with lower sediment transporting power.

Church, Annual Reviews of Earth Science, 2005

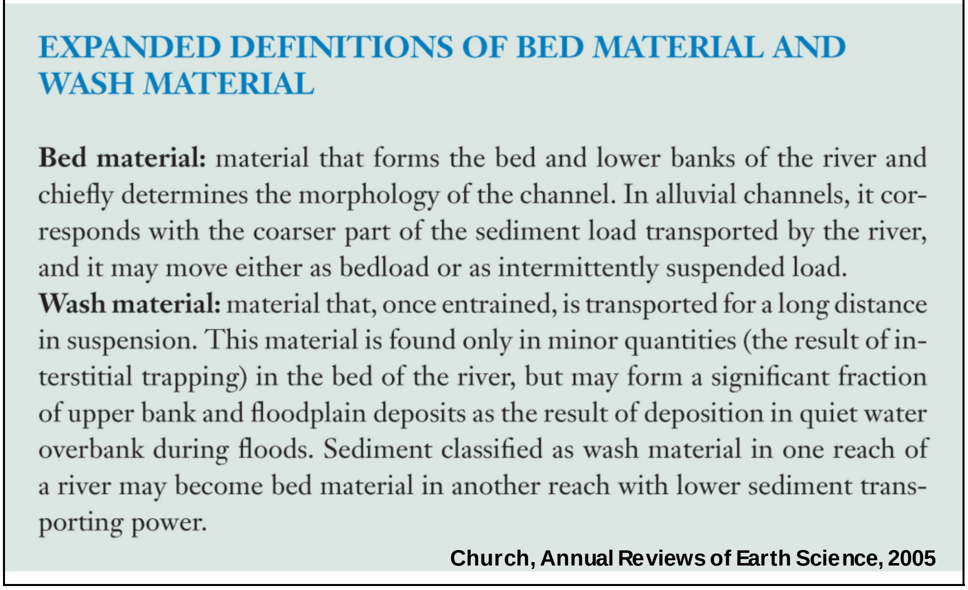
Goals of Rivers Lectures

1. How do landscape materials get from valley floors to their ultimate sink (oceans or lakes)?
2. Transition from hillslopes to river channels
   - i) Drainage basins and river networks
   - ii) Long valley profile and downstream fining
3. River channel morphology
4. River processes
   - i) Flow in rivers
   - ii) Sediment transport in rivers
   - iii) Bedrock incision by rivers

Sediment Transport Mechanisms



Sediment Transport in Rivers



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Church, Annual Reviews of Earth Science, 2005

Predicting Sediment Transport in Rivers

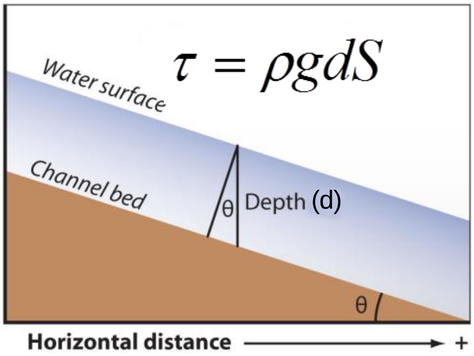
- We can predict bed material flux in rivers from an understanding of the physics of fluid flow.

- This is because bed material flux is driven by the force applied to the river bed (boundary shear stress).

- Predicting bed material sediment flux involves:

  1. Calculating the shear stress
  2. Establishing the entrainment and suspension thresholds
  3. Predicting bedload transport
  4. Predicting suspended sediment transport

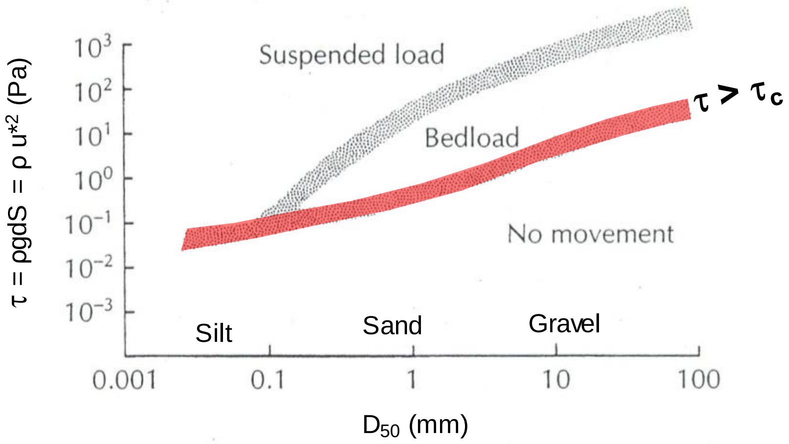
Calculation of the fluid shear stress



The **shear stress** (τ) exerted on the channel bed by the flow is equal to the downslope component of the weight of the overlying water $\tau = \rho_w g d S$, where ρ_w is the density of water and g is the acceleration due to gravity. The small angle approximation, where $S \sim \sin\theta$, is often used.

Bierman and Montgomery, 2014

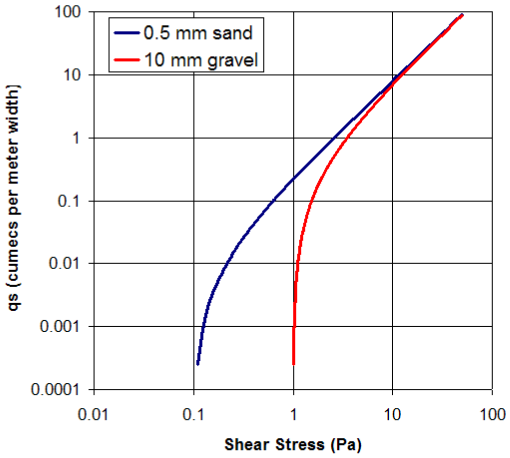
The Shields Curve



To establish the bedload entrainment threshold, we use the Shields diagram to find (τ_c). Then we determine if the fluid shear stress $\tau > \tau_c$. If so, bedload transport occurs.

Predicting bedload transport in rivers

We use a bedload transport equation to predict the amount of sediment moved in traction or saltation.

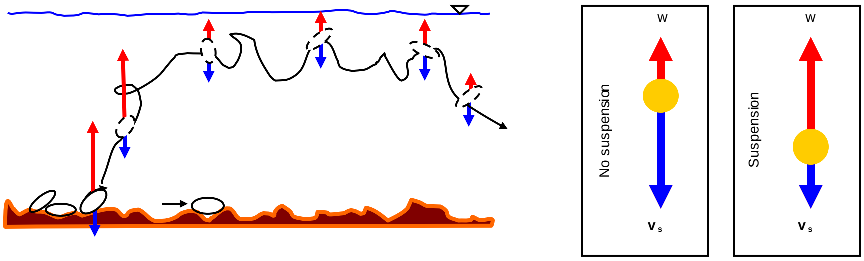


There are dozens of bedload equations, but the simplest one (Meyer-Peter and Muller Equation) takes the form:

$$q_s = 0.253(\tau - \tau_c)^{1.5}$$

q_s = sediment flux (sediment transport rate or sediment discharge) per unit width of the river [L^2/T]

The suspension threshold

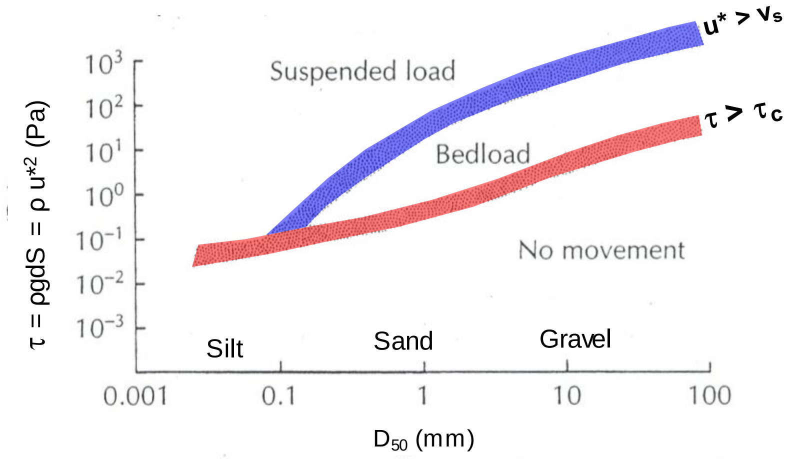


In order to move a particle into suspension, there must be a vertical fluid motion (w) away from the bed that is greater than the settling velocity of the sediment grain (v_s). These vertical fluid motions come from turbulence.

Research has shown that the vertical motions responsible for sediment suspension can be characterized by something called the shear velocity which is calculated from the shear stress as:

$$u* = \sqrt{\tau / \rho_w}$$

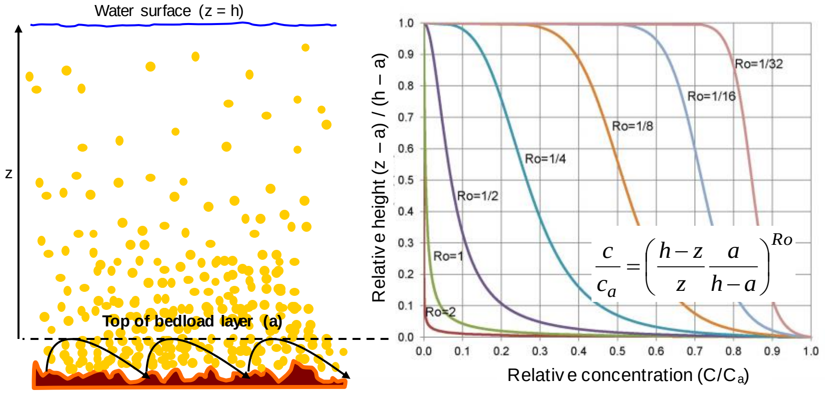
The suspension threshold on the Shields curve



We can capture this idea on the Shields by drawing a line where the shear velocity (u*) is greater than the settling velocity of sediment particles (v_s). If u* > v_s, the sediment will be entrained from the bedload into suspension.

Predicting suspended bed material load in rivers

In order to predict suspended bed material sediment flux in rivers, we use the Rouse equation that has the following form:

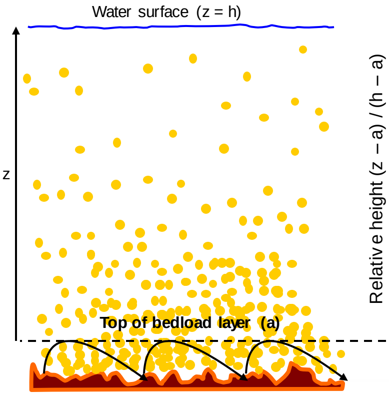


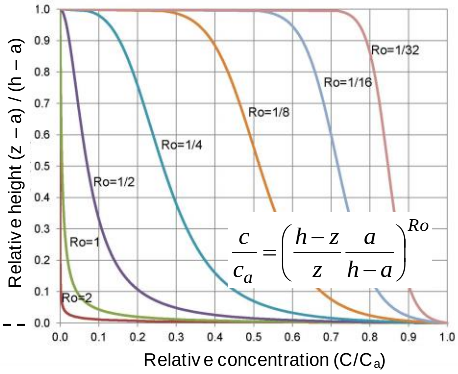
$$\frac{c}{c_a} = \left(\frac{h-z}{z} \frac{a}{h-a} \right)^{Ro}$$

The Rouse equation gives us the sediment concentration in the water column relative to the sediment concentration at the top of the bedload layer.

Predicting suspended bed material load in rivers

In order to predict suspended bed material sediment flux in rivers, we use the Rouse equation that has the following form:

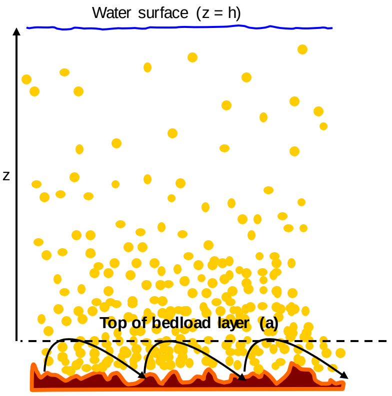




$$\frac{c}{c_a} = \left(\frac{h-z}{z}\,\frac{a}{h-a}\right)^{Ro}$$

Ro is the Rouse number (~v_s / u*). When it is large, v_s >> u*, so there is no suspension (Ro > 2.5). When it is small, v_s < u*, so there is lots of suspension high up in the flow (Ro < 1).

Predicting bed material load in rivers

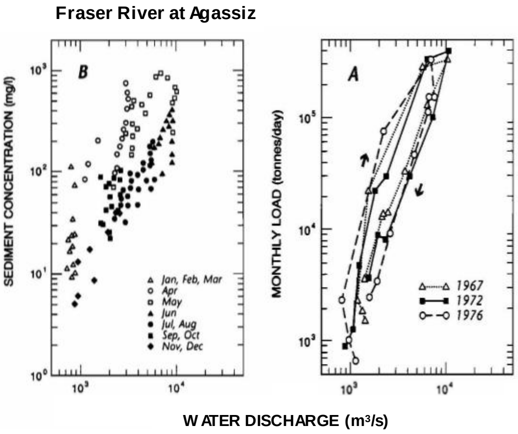


By multiplying the concentration at any height above the bed by a velocity and a depth increment, we get sediment flux at a given height above the bed. Summing up vertically, we get the suspended bed material flux (per unit width of the river [L^2/T])

Total bed material sediment flux (Q_s) in a cross-section of a river is the sum of the bedload and the suspended bed material sediment flux per unit width of the river times the channel width.

$$Q_s = (q_{s\text{-bedload}} + q_{s\text{-ssbm}}) * \text{width}$$

Predicting washload in rivers



Unfortunately, we cannot predict wash material load in rivers from physics because it is controlled by sediment supply from the hillslopes and not the capacity of the channel to carry sediment. So geomorphologists usually rely on suspended sediment rating curves.

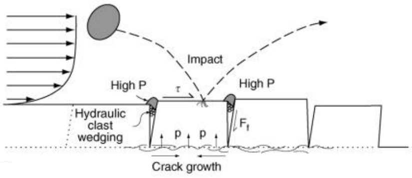
$$q_s = Cq$$

q = water flux per unit width of the river [L^2/T]
C = sediment concentration in water column [dimensionless]

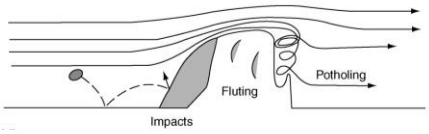
Sediment transport in rivers

Bedrock Erosion Processes (in rivers)

Plucking



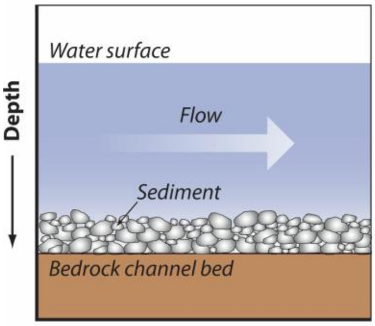
Abrasion



Whipple, DiBiase and Crosby, 2013

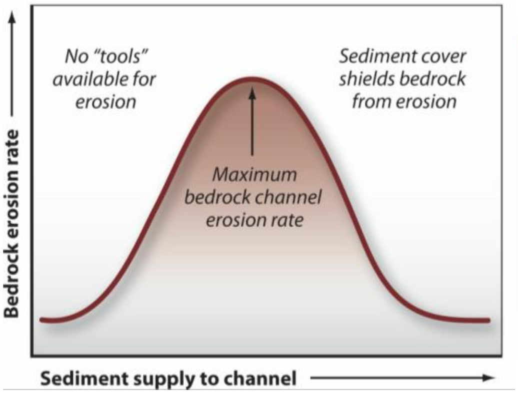
Tools and Cover Effect

River channels are underlain by both rock and sediment. In order for the river to incise into the underlying bedrock, it must first entrain sediment from the channel bed. Bedrock cannot be eroded until its sediment cover is in motion.



Bierman and Montgomery, 2014

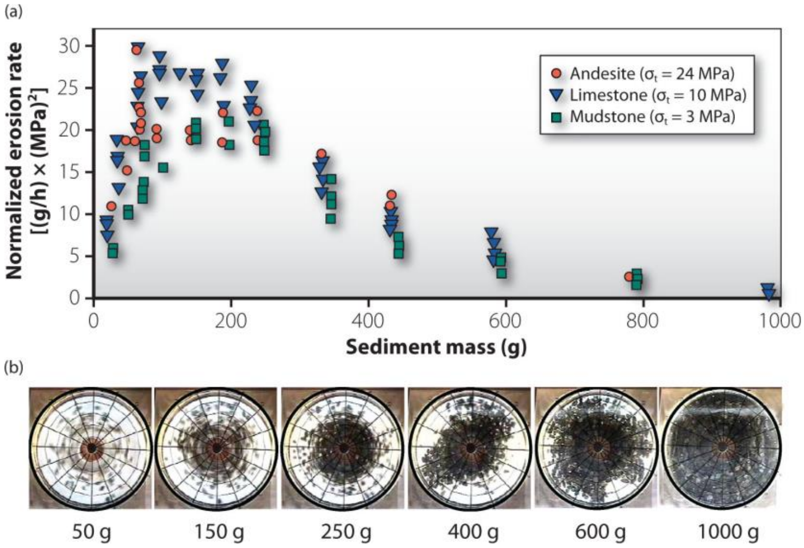
Tools and Cover effect



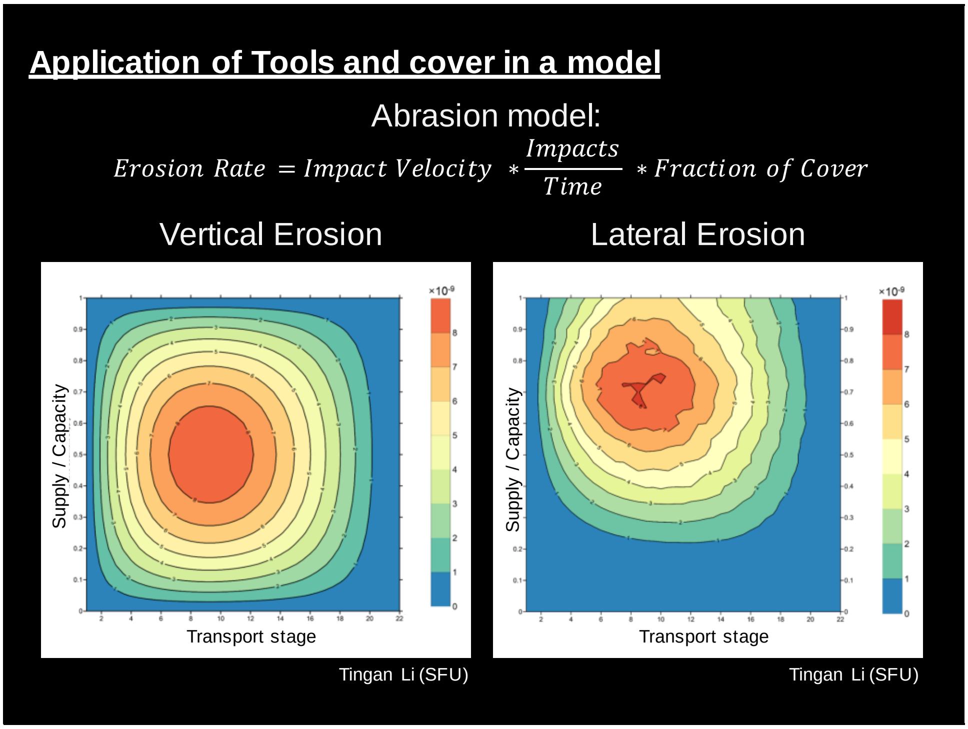
The rate of **bedrock incision** also depends on the sediment supply to the channel. The maximum bedrock erosion rate occurs when there are enough "tools" (clasts) entrained in the flow to erode the bed, but not enough sedimentary cover to protect the underlying bedrock from erosion.

Bierman and Montgomery, 2014

Tumbler experiments Sklar and Dietrich (2001)



Bierman and Montgomery, 2014



Application of Tools and cover in a model

Abrasion model:

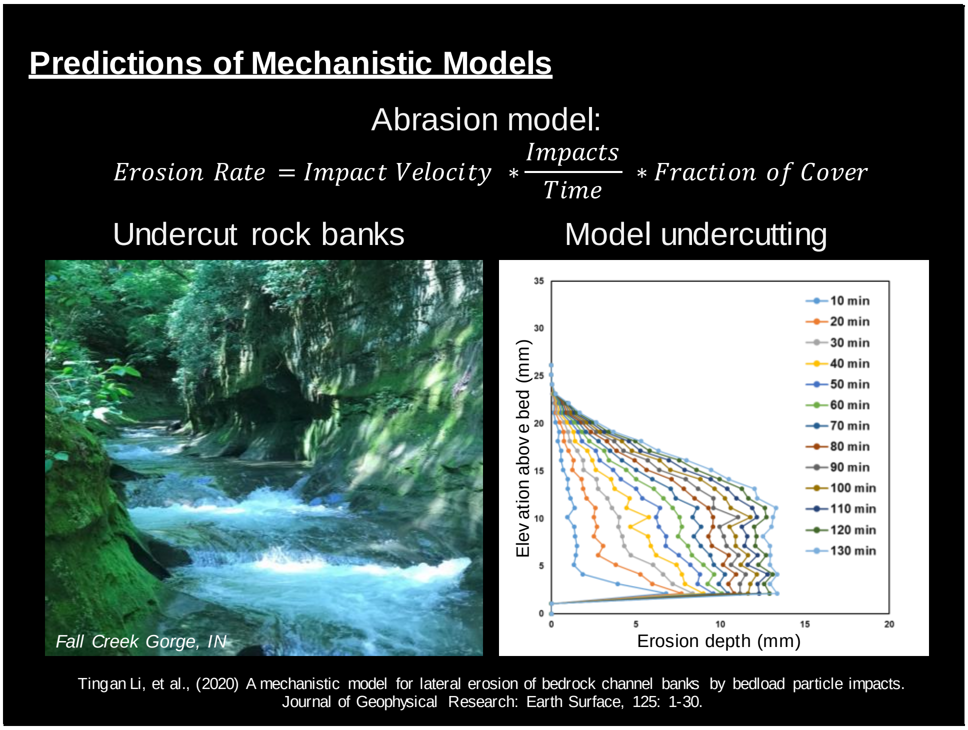
$$Erosion\ Rate\ = Impact\ Velocity\ * \frac{Impacts}{Time} * Fraction\ of\ Cover$$

Vertical Erosion

Lateral Erosion

Tingan Li (SFU)

Tingan Li (SFU)



Predictions of Mechanistic Models

Abrasion model:

$$Erosion\ Rate\ = Impact\ Velocity\ * \frac{Impacts}{Time} * Fraction\ of\ Cover$$

Undercut rock banks

Model undercutting

Fall Creek Gorge, IN

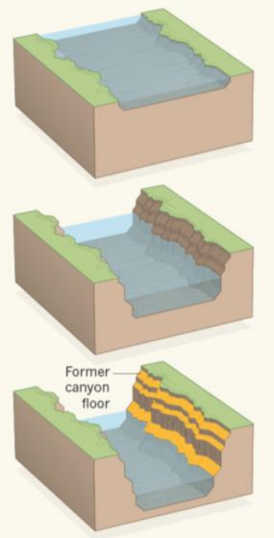
Tingan Li, et al., (2020) A mechanistic model for lateral erosion of bedrock channel banks by bedload particle impacts. Journal of Geophysical Research: Earth Surface, 125: 1-30.

Reach-scale predictions of erosion

Progressive erosion by plucking of columnar basalt

Plucking model: $E \propto \left(\tau - \tau_{pc}\right)^{3/2}$

Moses Coulee, Scablands, Eastern WA

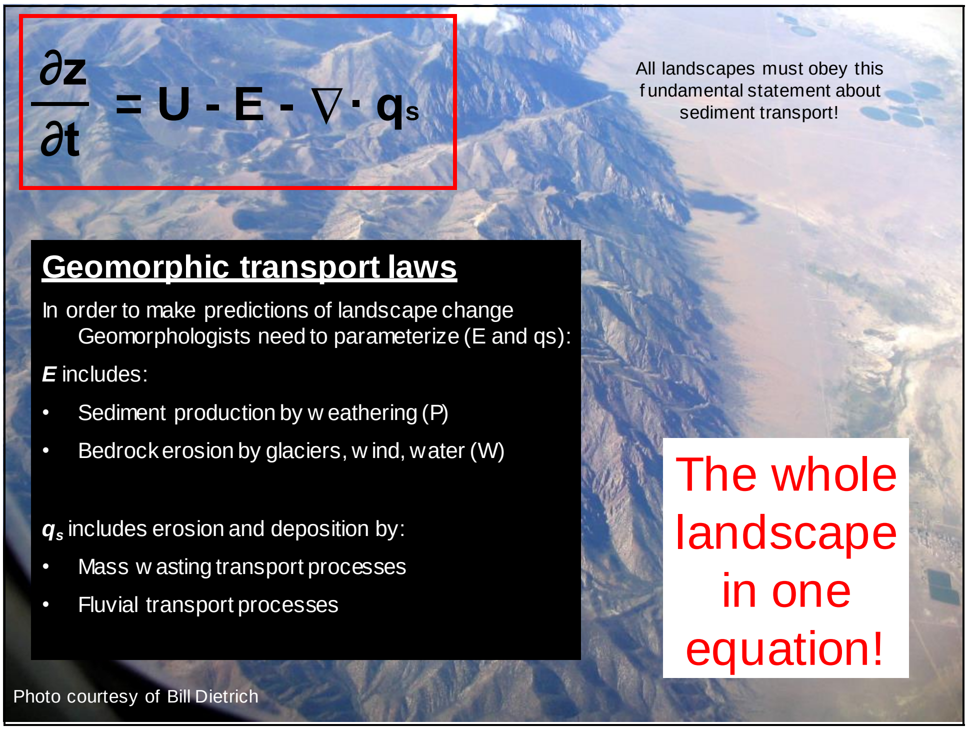


Cartoon from Perron and Venditti (2016)
Model by Larsen and Lamb (2016)

Progressive incision of the Channeled Scablands by outburst floods

Isaac J. Larsen[1,2] & Michael P. Lamb[2]

doi:10.1038/nature19817



$$\frac{\partial z}{\partial t} = U - E - \nabla \cdot q_s$$

All landscapes must obey this fundamental statement about sediment transport!

Geomorphic transport laws

In order to make predictions of landscape change Geomorphologists need to parameterize (E and qs):

E includes:

- Sediment production by weathering (P)
- Bedrock erosion by glaciers, wind, water (W)

$\boldsymbol{q_s}$ includes erosion and deposition by:

- Mass wasting transport processes
- Fluvial transport processes

The whole landscape in one equation!

Photo courtesy of Bill Dietrich