Earth-surface Dynamics Modeling & Model Coupling

A short course

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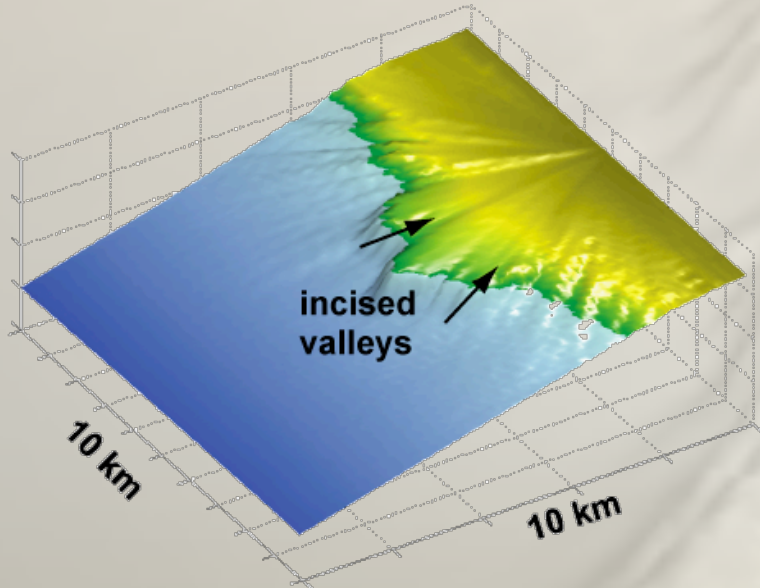
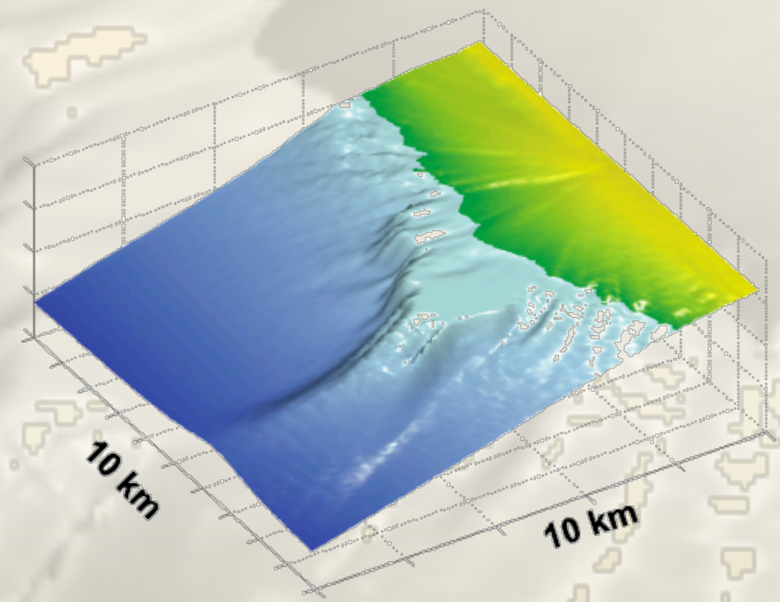
Special thanks to Alan Howard, Sergio Garcia, Irina Overeem, Scott Peckham





COMMUNITY SURFACE DYNAMICS MODELING SYSTEM

Module 4: Coastal Morphodynamics

ref: Hutton E.W.H., and Syvitski, J.P.M. 2008, SedFlux2.0: New advances in the seafloor evolution and stratigraphic modular modeling system. Computers and Geosciences. 34, 1319-1337.

Relative Sea Level (5)

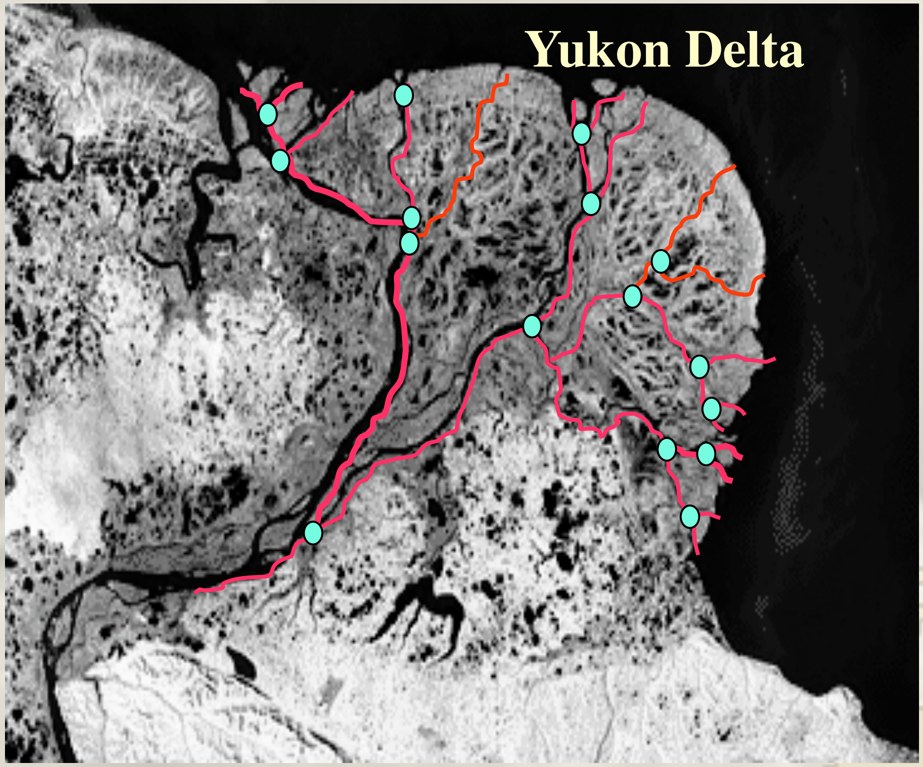
Drainage Basins & Sea Level (7)

Barrier Islands & Sea Level (4)

Deltas & Channel Switching (13)

Alongshore & Cross-Shore Transport Modeling (6)

Summary (1)



Controls on Surface Elevation Above Mean Sea Level

$$D_{RSL} = A - \Delta E - C_n - C_A \pm M$$

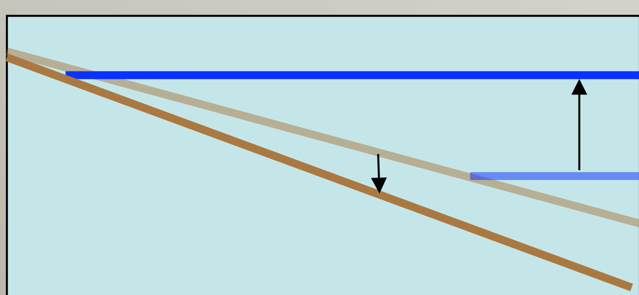
•D_{RSL} = Elevation change relative to mean sea level
•A = *Aggradation Rate:* sediment delivered to and retained on the subaerial surface as new sedimentary layers
•ΔE = *Eustatic Sea Level Rate:* changes to the volume of the global ocean over time, as influenced by fluctuations in the storage of terrestrial water (glaciers, ice sheets, groundwater, lakes, reservoirs), and ocean water expansion due to T°C changes
•C_n = *Natural Compaction:* natural changes in the void space within sedimentary layers (dewatering, grain-packing realignment, organic matter oxidation)
•C_A = *Accelerated Compaction*: anthropogenic contribution to volume change as a consequence of subsurface mining (oil, gas or groundwater), human-influenced soil drainage and accelerated oxidation
•M = vertical movement of the land surface as influenced by the redistribution of earth masses (e.g. sea level fluctuations, growth of deposits, growth/shrinkage of nearby ice masses, tectonics, deep-seated thermal subsidence).

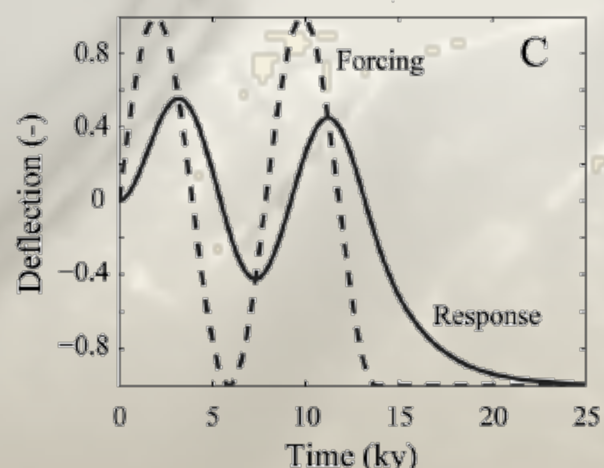
Geological Controls on the Relative Sea Level of a Delta

- **Isostasy** involves significant load changes to a regions crust
  - i) Growth/shrinkage of large ice masses (glacio-isostasy)
  - ii) Thick sediment deposits (sediment load isostasy)
  - iii) Water added/subtracted with fluctuations in sea level (hydro-isostasy)
- **Fault-controlled tectonics** can raise or lower a delta's surface
- **Thermal subsidence** — deep long-term response of passive margins

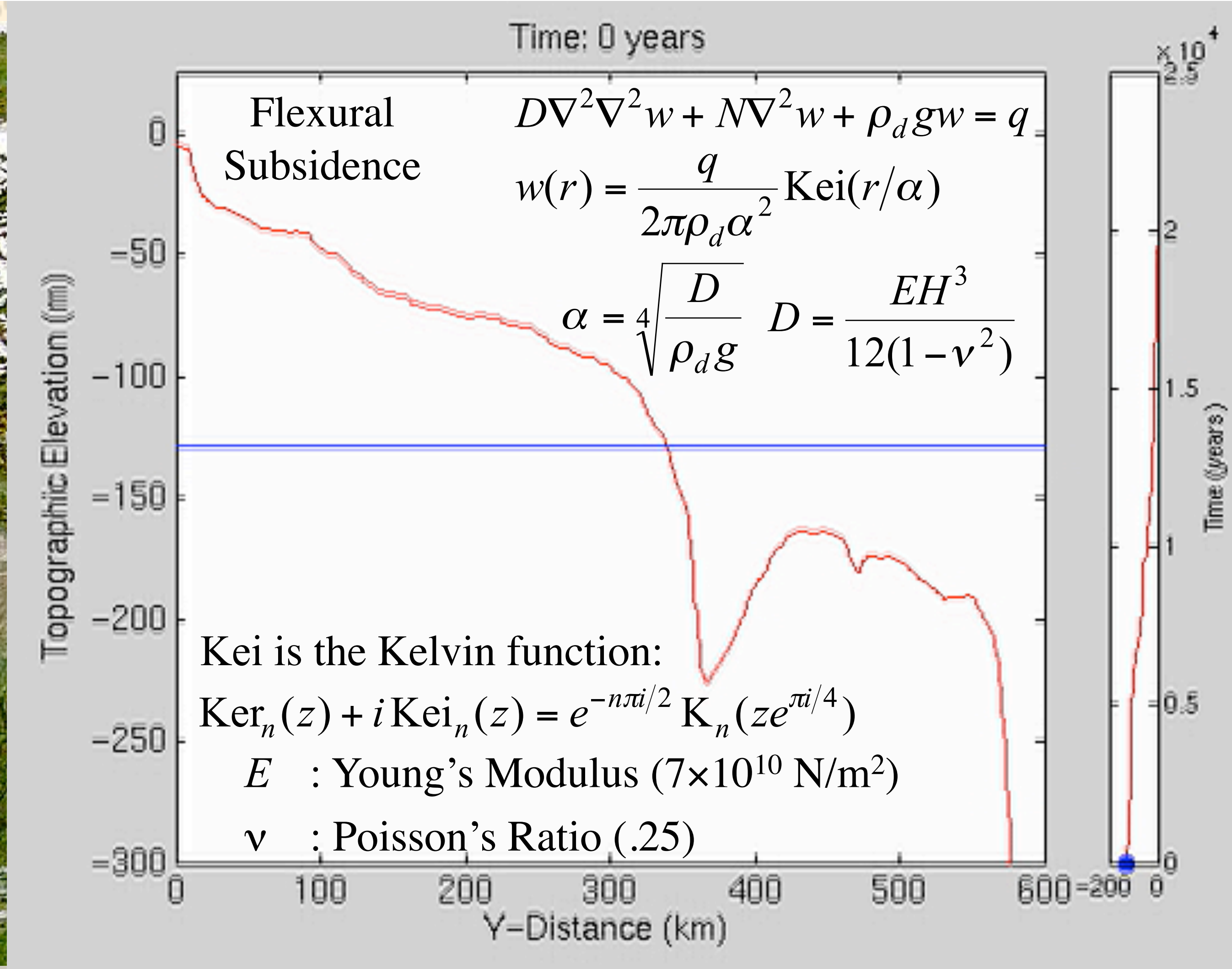
The crust takes thousands of years to relax (Flexural Response) to changes in load because the viscous asthenosphere has to flow out of the way before the lithosphere can deflect. E-folding time of this response is ≈2500y.

Isostatic displacements extend over a region much larger than the area directly affected by the load change (regional elastic lithosphere thickness).

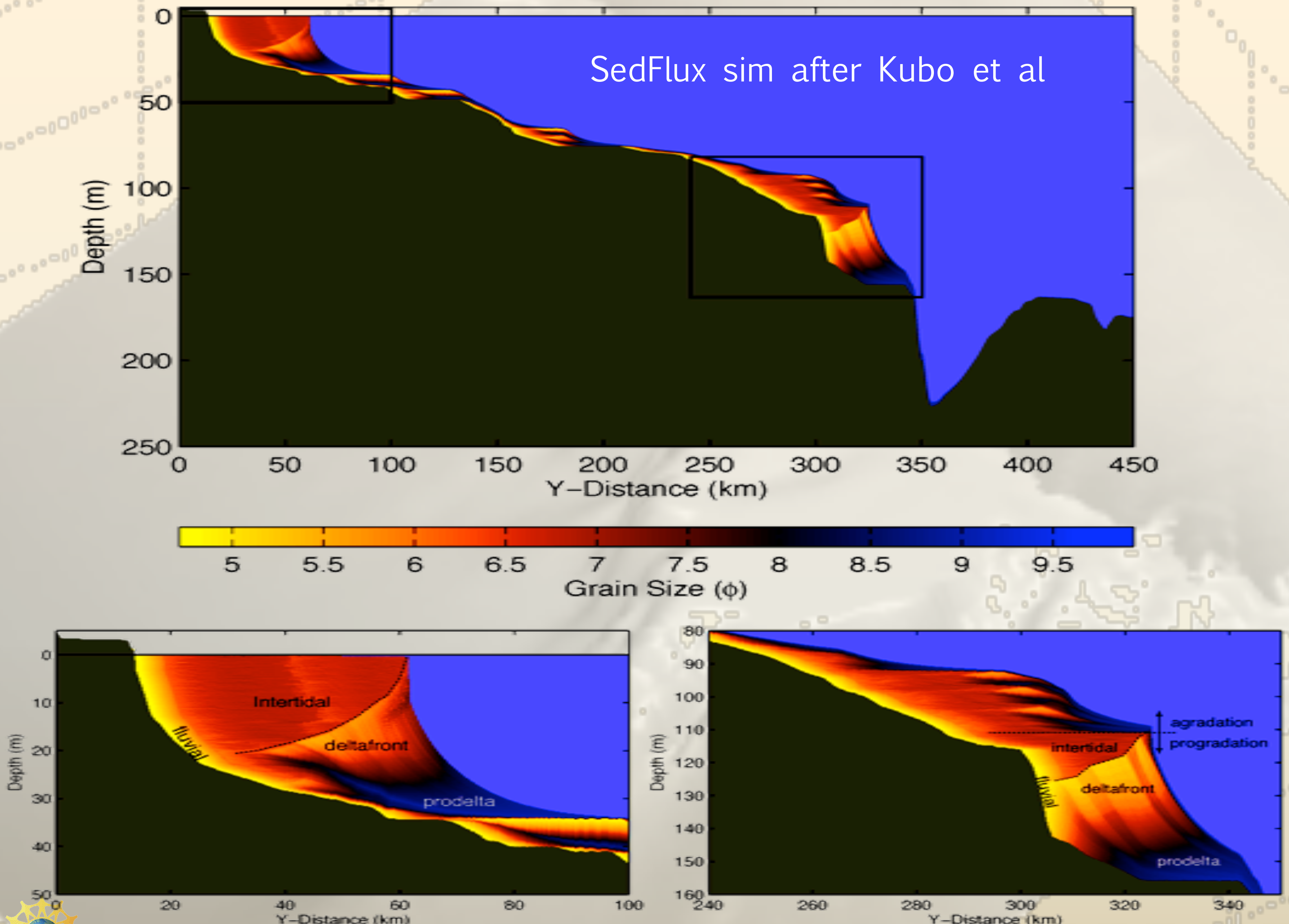




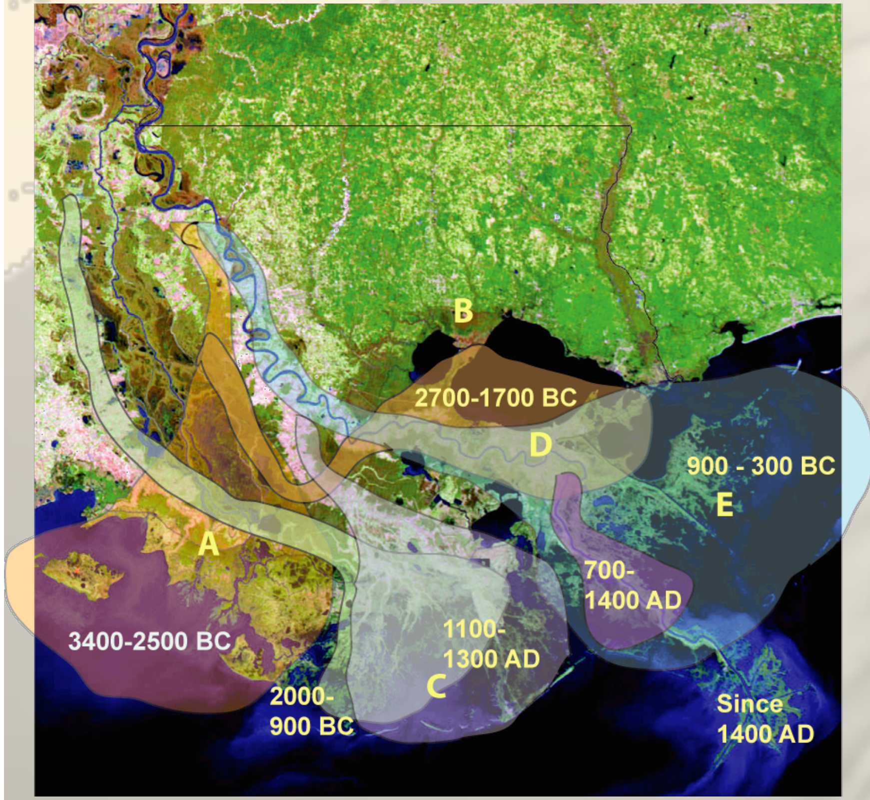
Holocene SL rise (hydro-isostasy) is still affecting the world's coastlines.

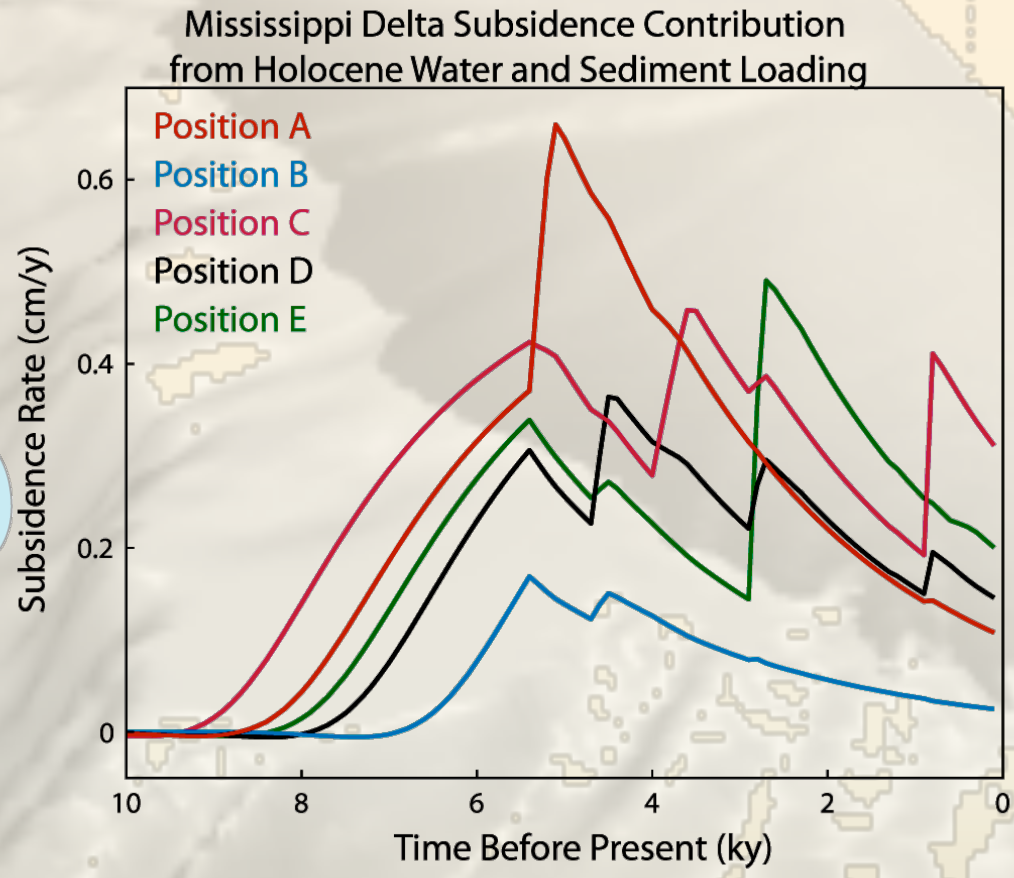


Numerical solution to the water load related to sea level rise in the Adriatic since the Last Glacial Maximum (21 Kyr) plus 10 Kyr into the future. Time step is 100 yr.


SedFlux sim after Kubo et al
Depth (m)
Y–Distance (km)
Grain Size (φ)
Intertidal
fluvial
deltafront
prodelta
agradation
progradation
intertidal

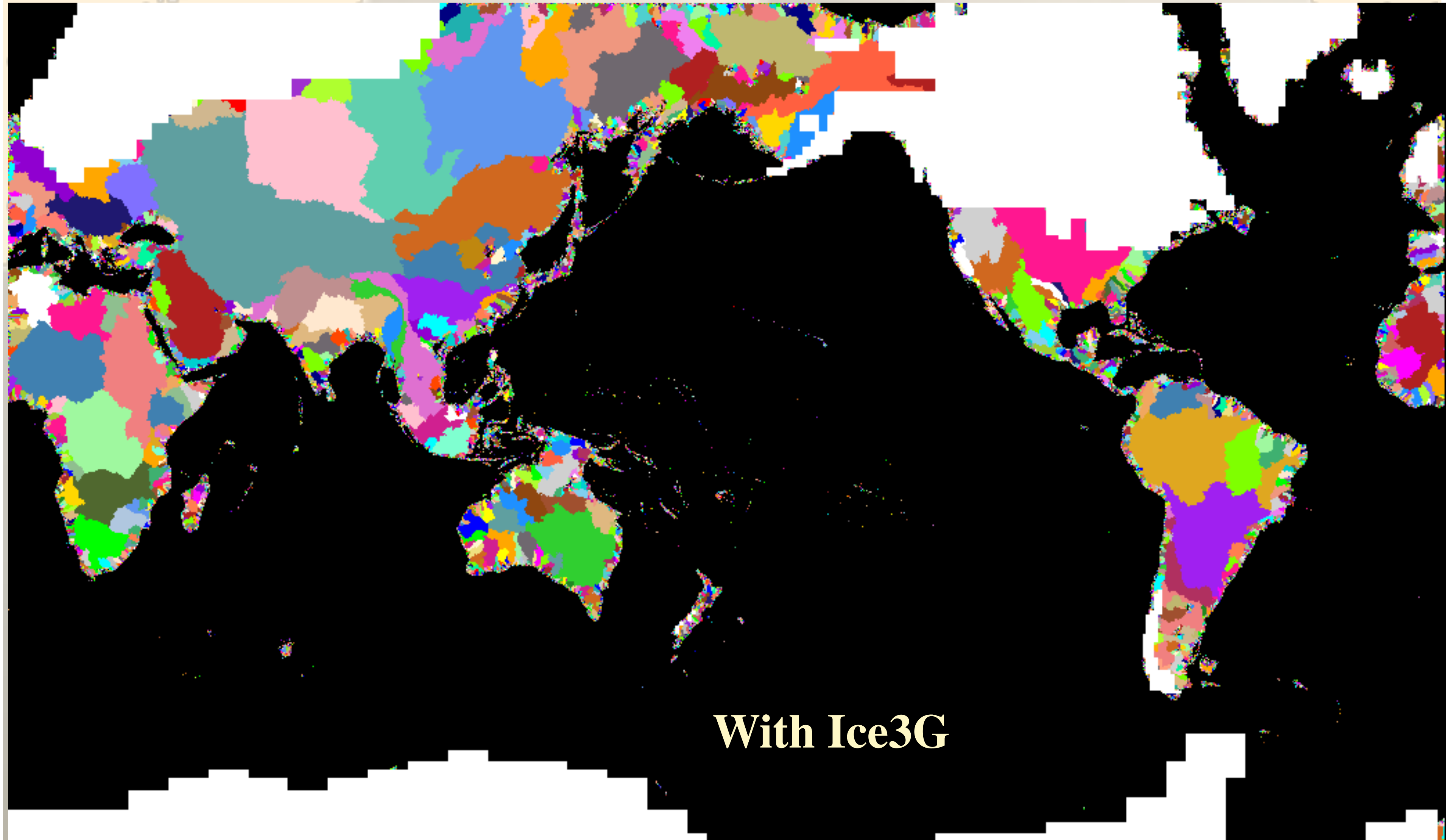
Geological Controls on the Relative Sea Level of a Delta

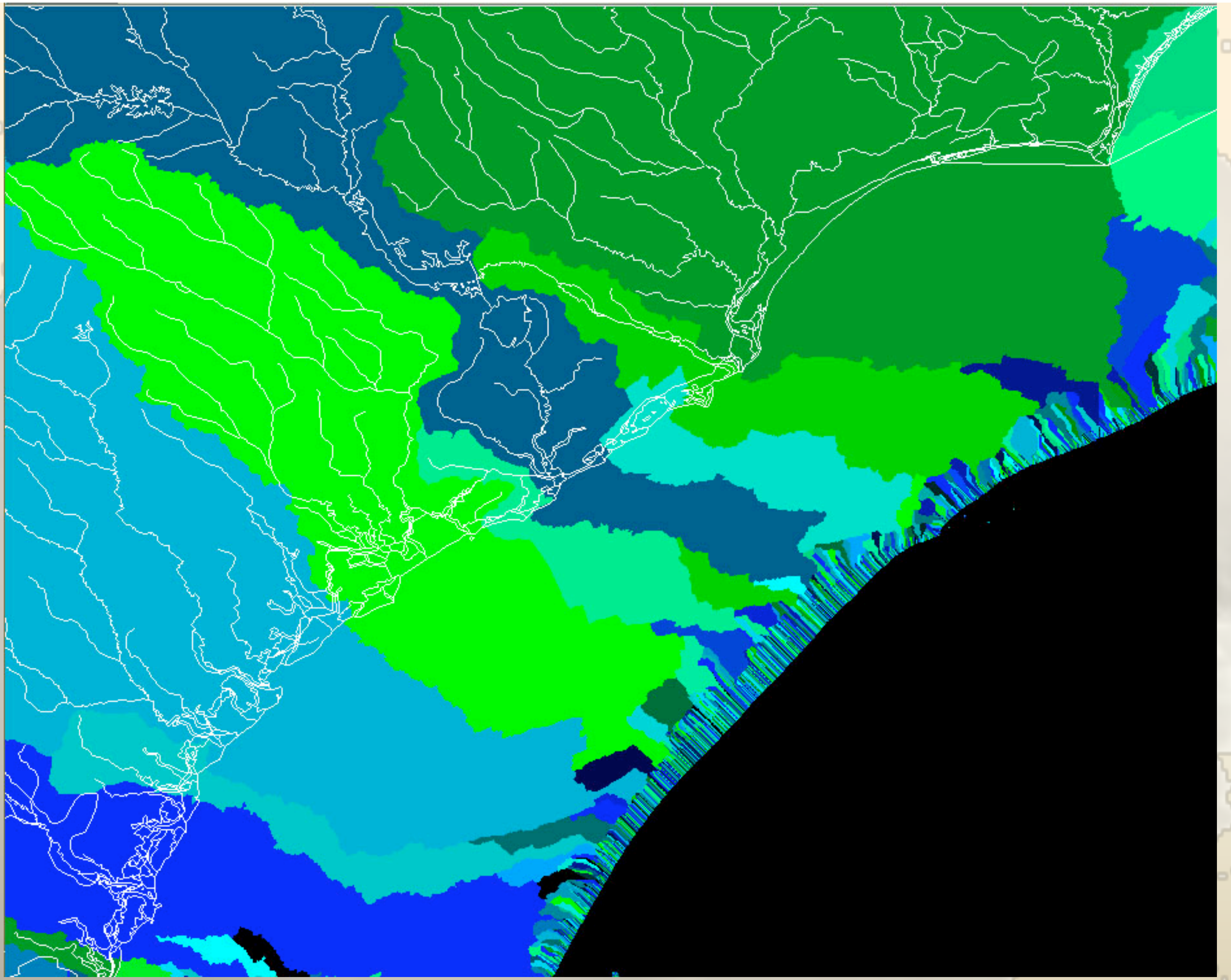


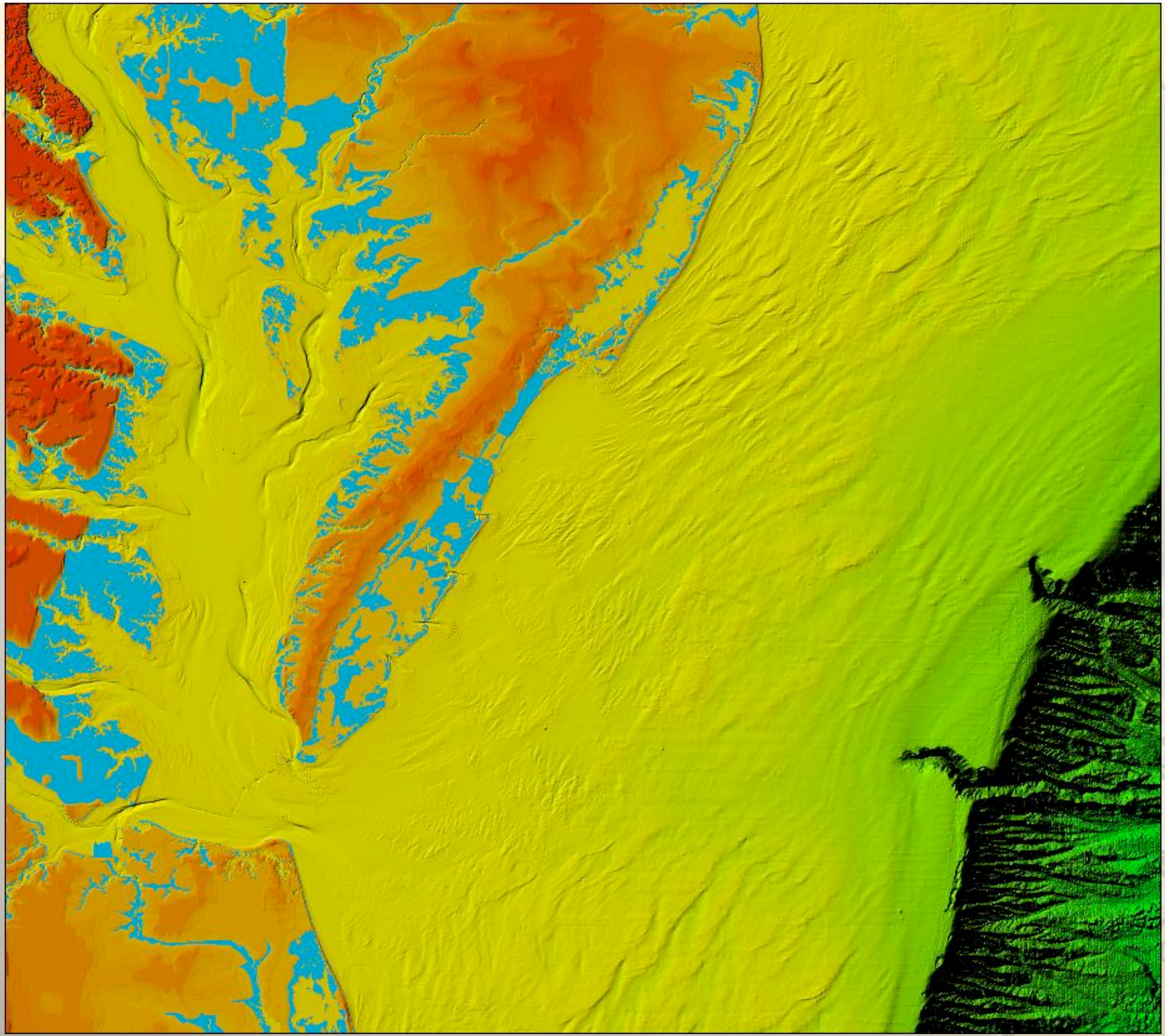


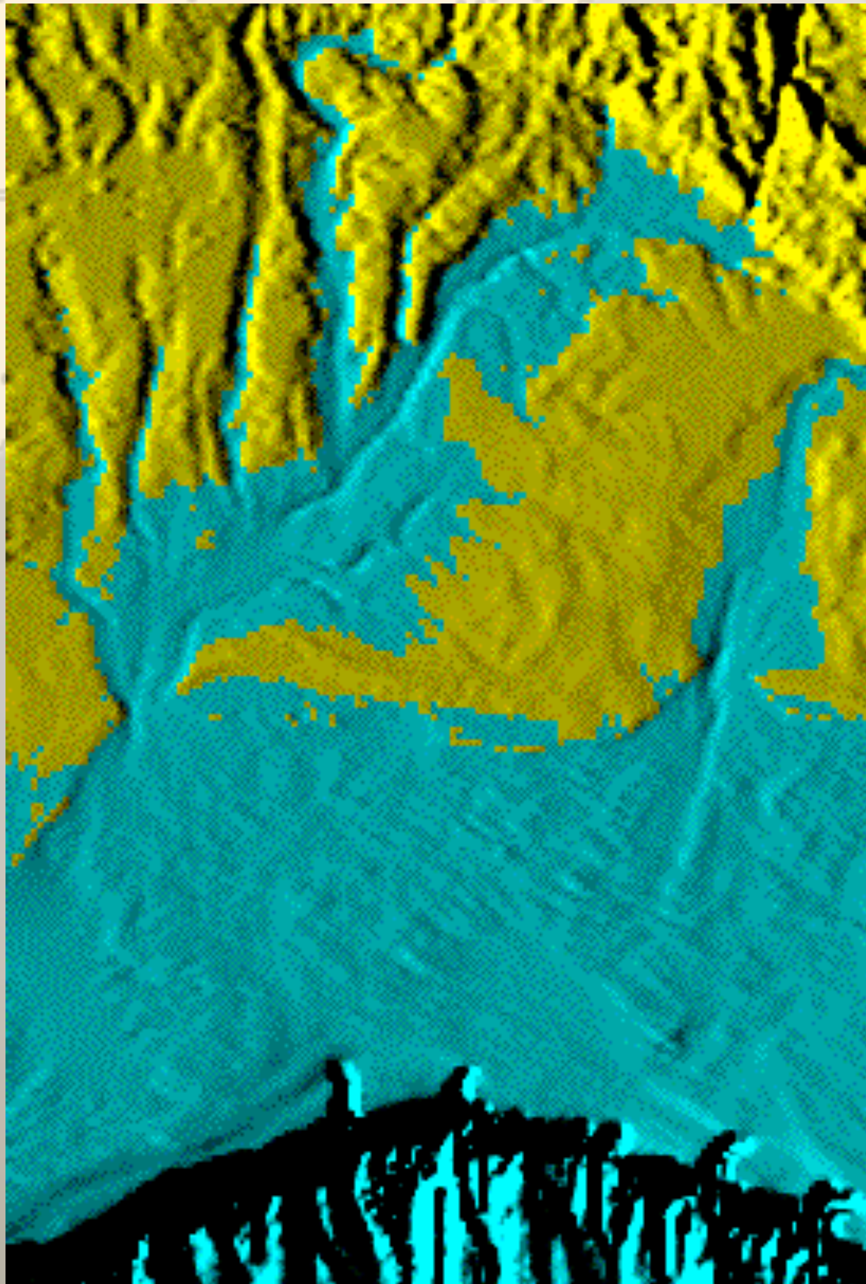
The various Mississippi delta lobes weigh between 200 to 900 billion tonnes. Each location on a large delta sinks at different rates, depending on their load history. Today the various Mississippi lobes are sinking at between 0.3 to 3.6 mm/y.

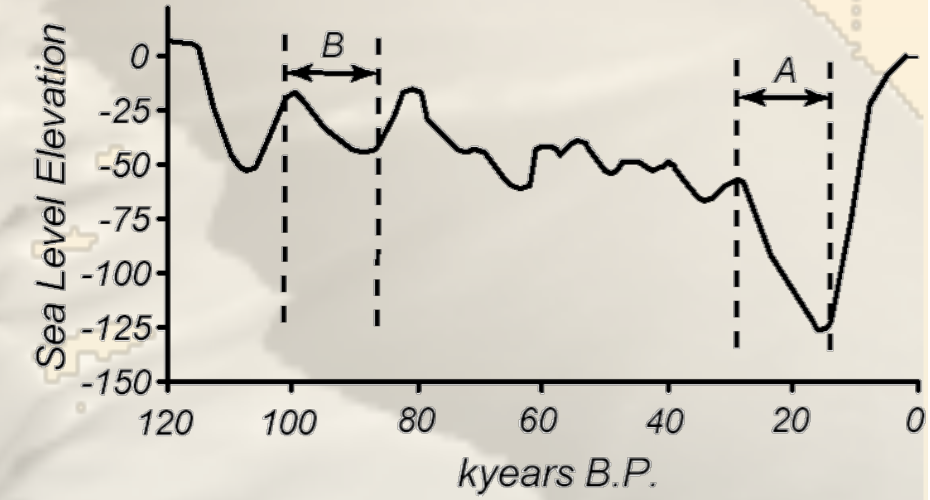
20ka Drainage Basins







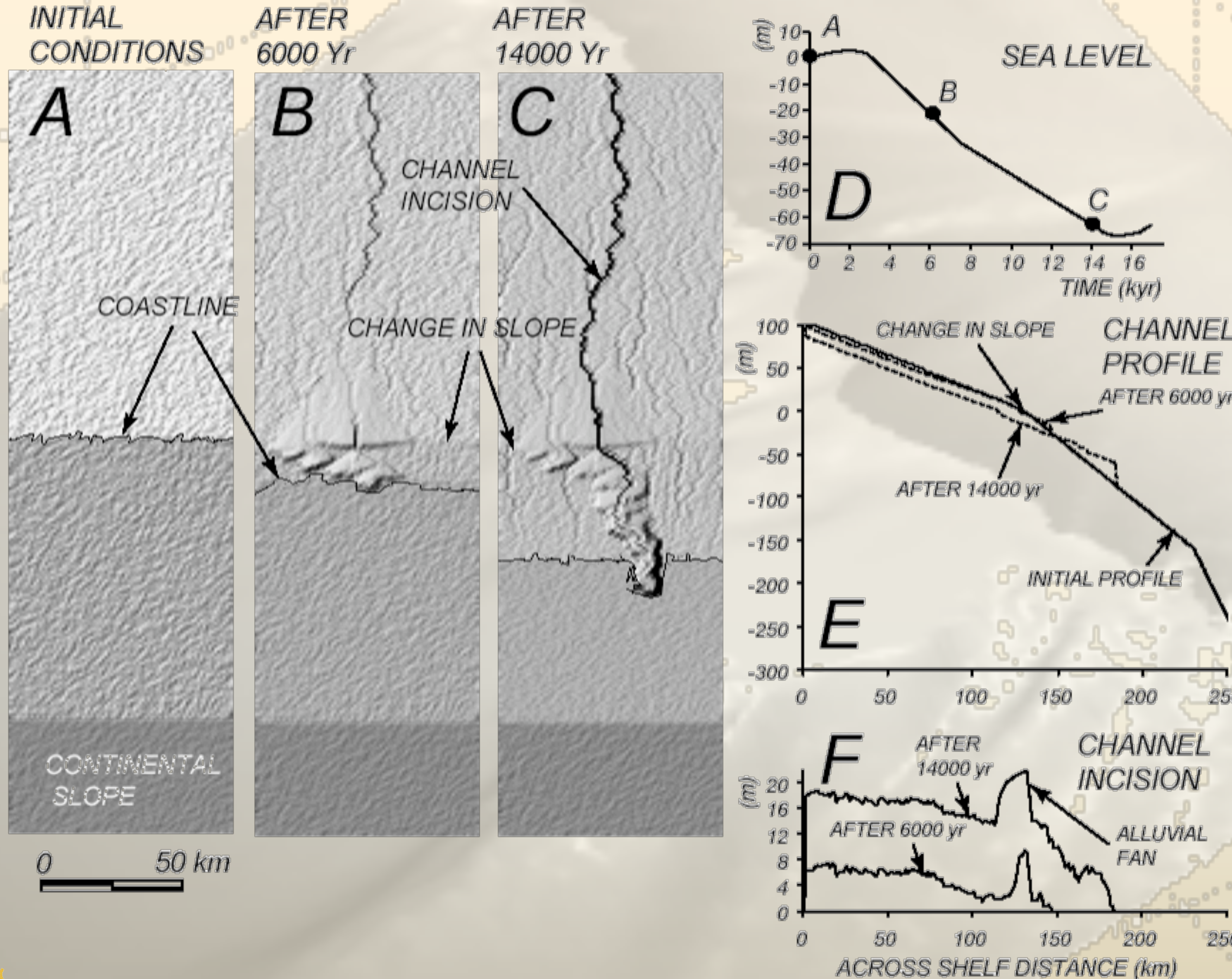




After Howard et al.

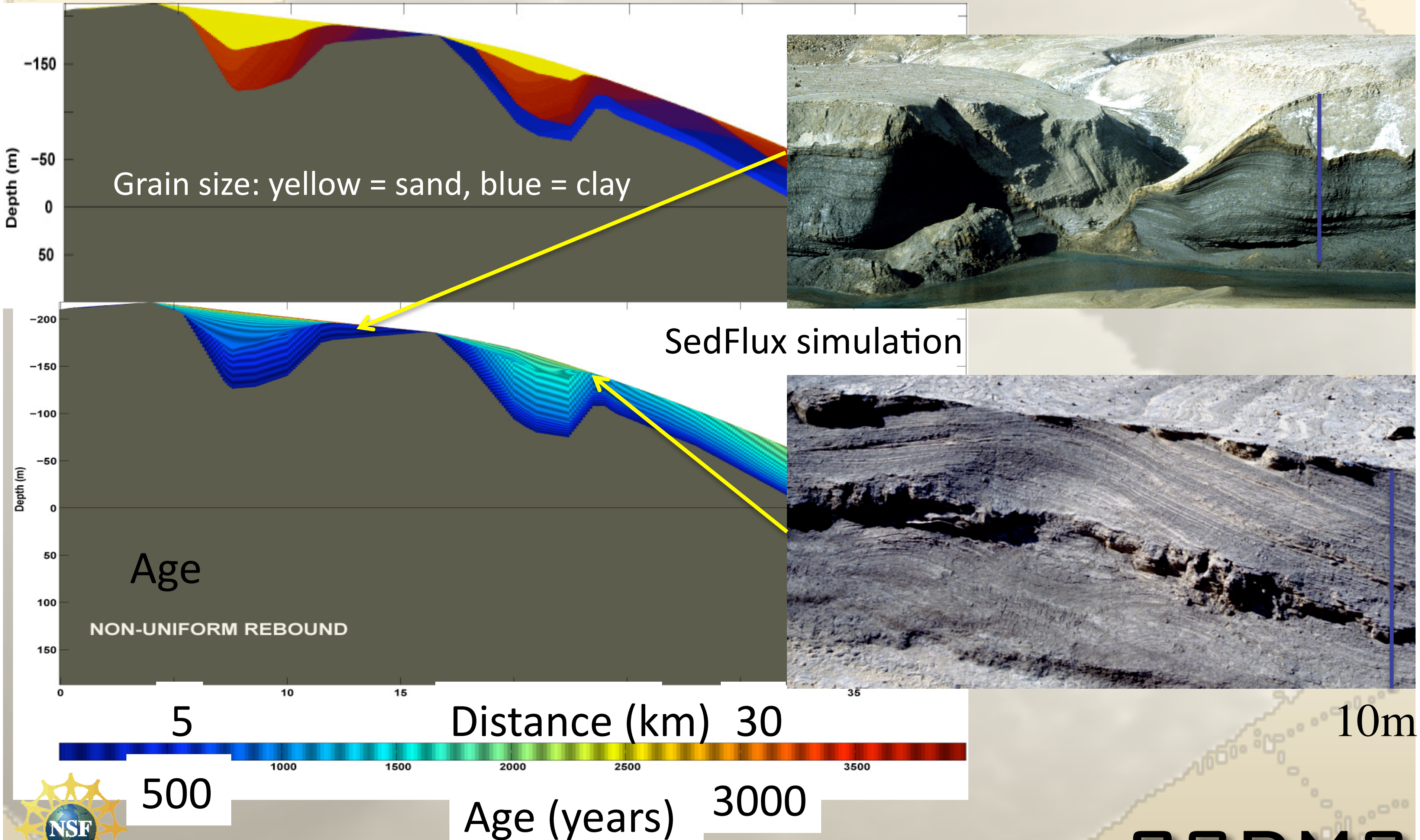
Observations from Simulations (after A. Howard)

- Initial topography strongly influences drainage patterns.
- Re-excavation of previous lowstand channel systems
- Most dissection occurs in a few major channels. Channels excavate as much as 20m in New Jersey simulation and 41m in Virginia simulation.
- The main drainage pattern is established very early in the simulation
- Depressions are infilled and eventually integrated into drainage system.



INITIAL CONDITIONS
AFTER 6000 Yr
AFTER 14000 Yr
A
B
C
CHANNEL INCISION
COASTLINE
CHANGE IN SLOPE
CONTINENTAL SLOPE
0
50 km
SEA LEVEL
D
(m)
10
0
-10
-20
-30
-40
-50
-60
-70
0
2
4
6
8
10
12
14
16
TIME (kyr)
CHANGE IN SLOPE
CHANNEL PROFILE
AFTER 6000 yr
AFTER 14000 yr
INITIAL PROFILE
E
100
50
0
-50
-100
-150
-200
-250
-300
0
50
100
150
200
250
F
AFTER 14000 yr
CHANNEL INCISION
AFTER 6000 yr
ALLUVIAL FAN
20
16
12
8
4
0
ACROSS SHELF DISTANCE (km)

During falling SL, a clinoform emerges, local depocenters are filled in. Rapid local fluvial incision, razor-blade effect in zone of highest uplift



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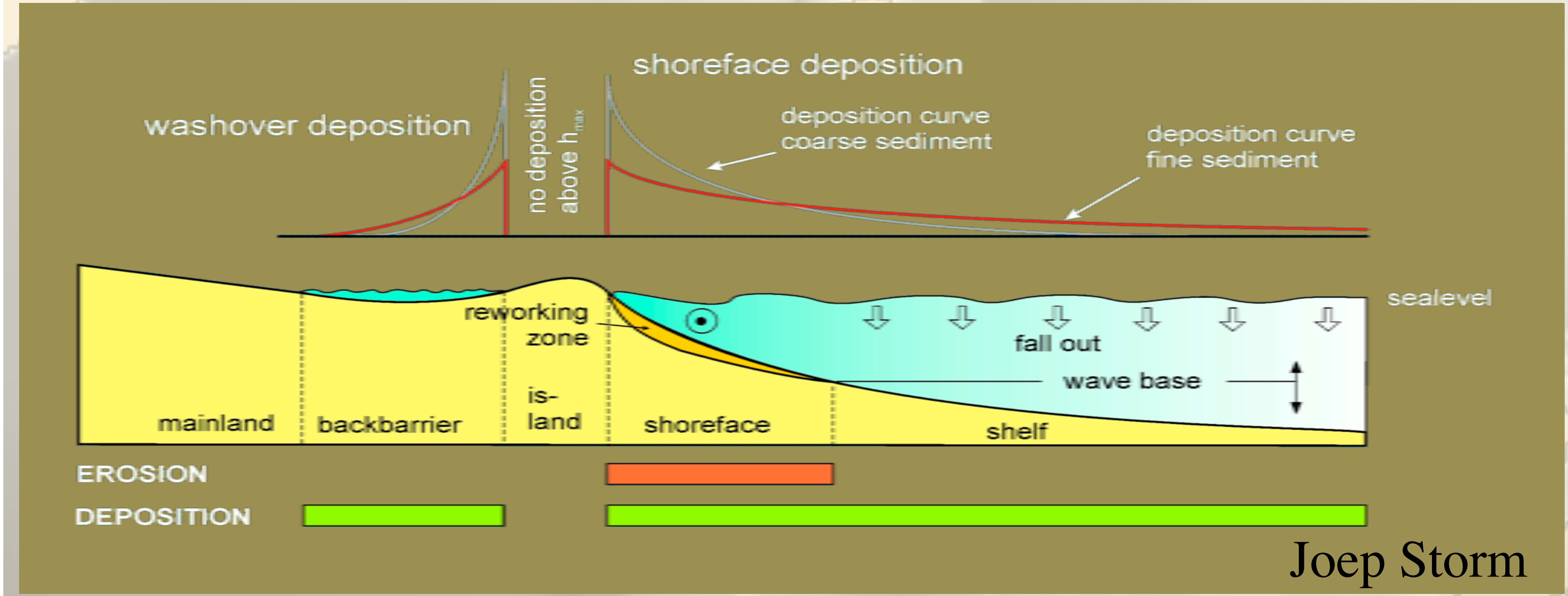
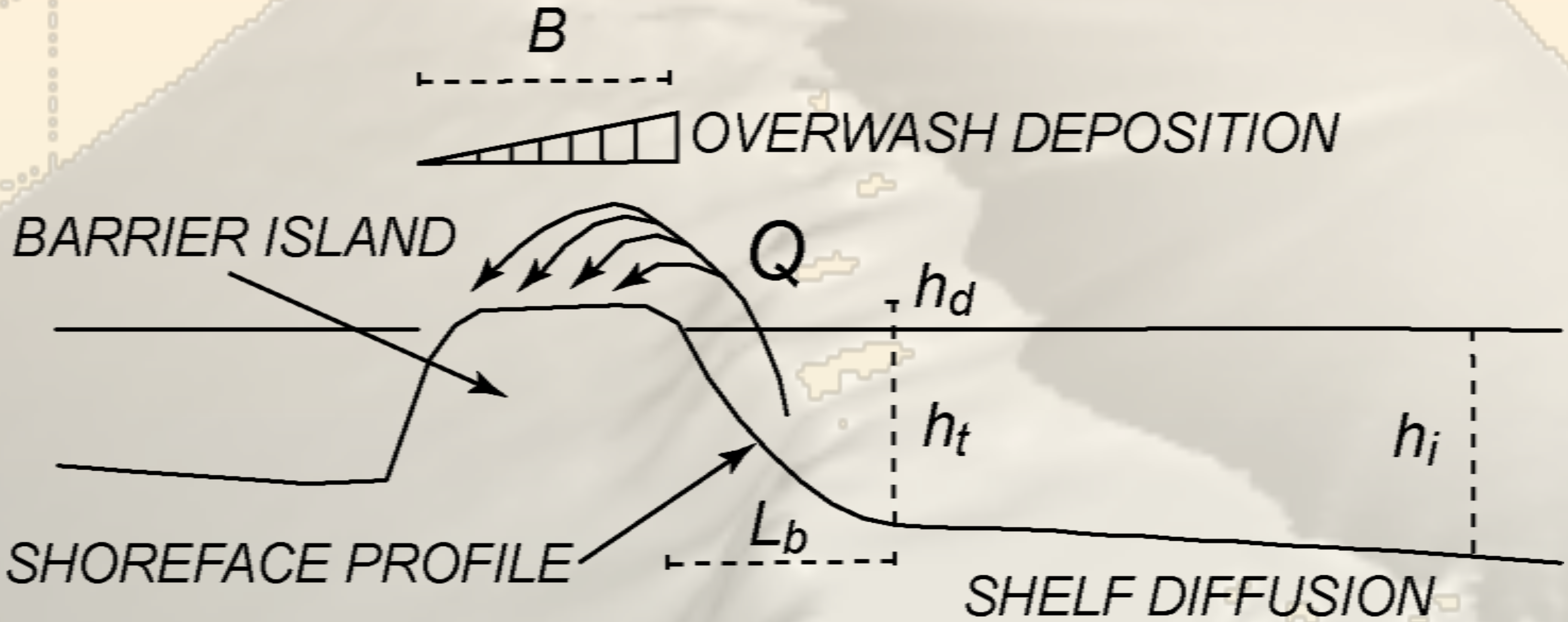
Modeling Shoreline Development

In landscape evolution models it is extremely important to keep track of the coastline position, since the beach profile is usually the steepest element of the inner shelf



Modeling Shoreline Development: Sergio Fagherazzi



To study barrier island evolution, consider three different processes with different timescales:

Formation of a shoreface equilibrium profile (fast)

$$\frac{\partial h}{\partial t} = K' \frac{\partial^2}{\partial x^2}\left(h - h_{eq}\right)$$

$$h_{eq} = a x^m$$

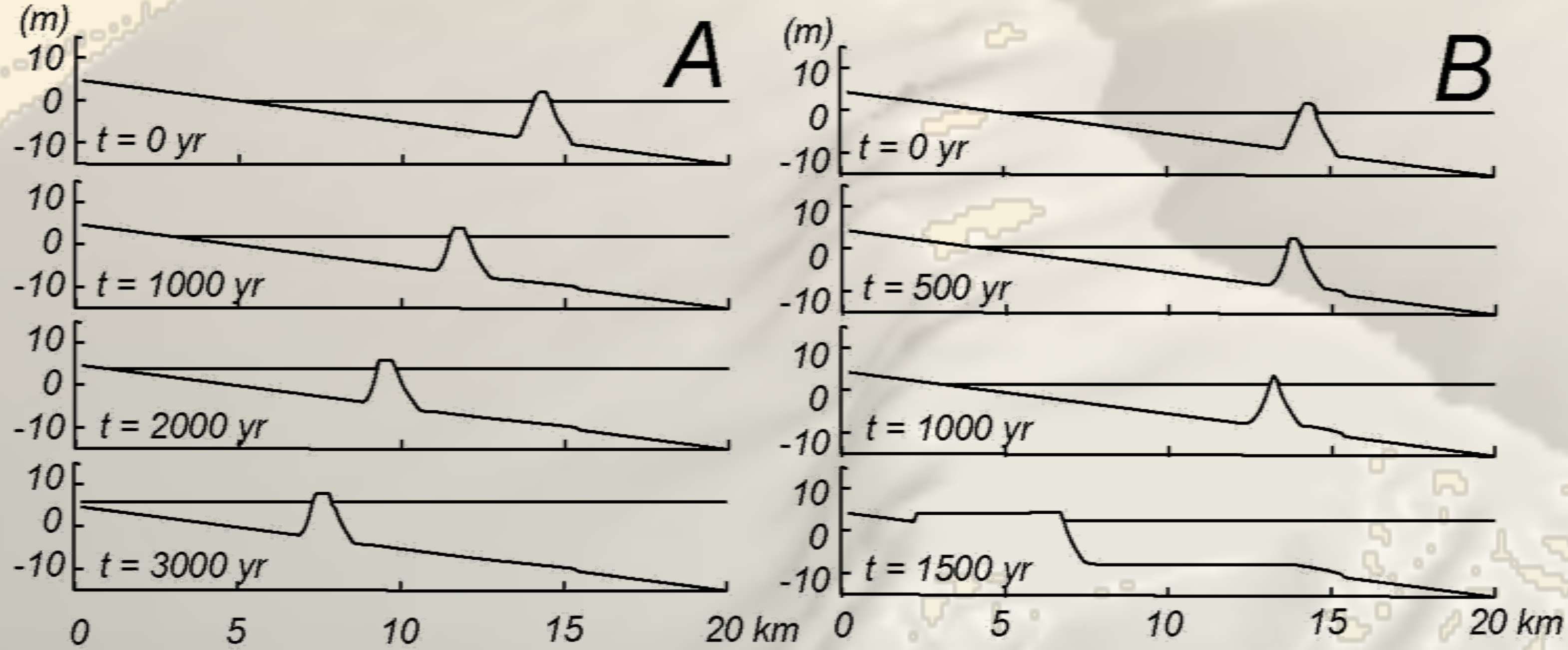
Inner shelf diffusion (slow)

Overwash (slow)

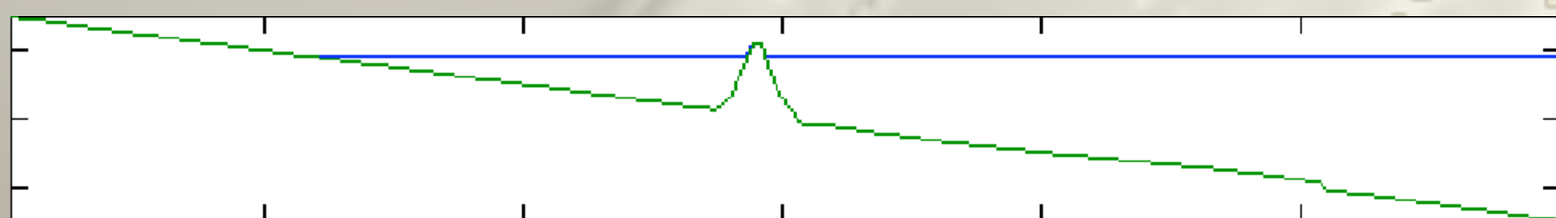
$$\frac{\partial h}{\partial t} = D\left(\frac{h_t - h}{h_t - h_i}\right)\frac{\partial^2 h}{\partial x^2}$$

Modeling Shoreline Development

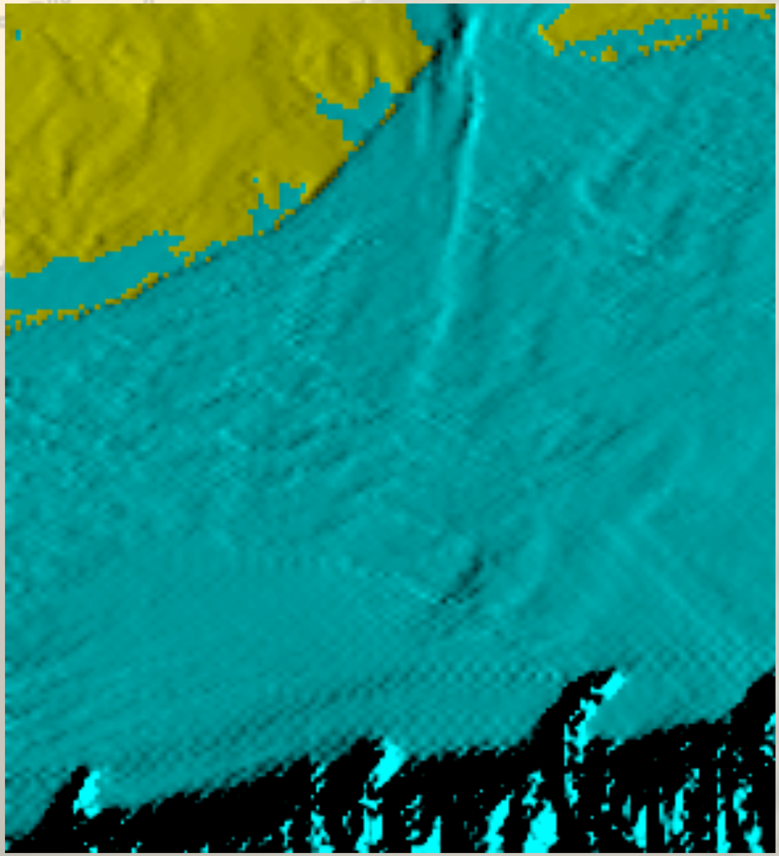
Comparison between the two simulations:
A drowning barrier island translates at a slower rate



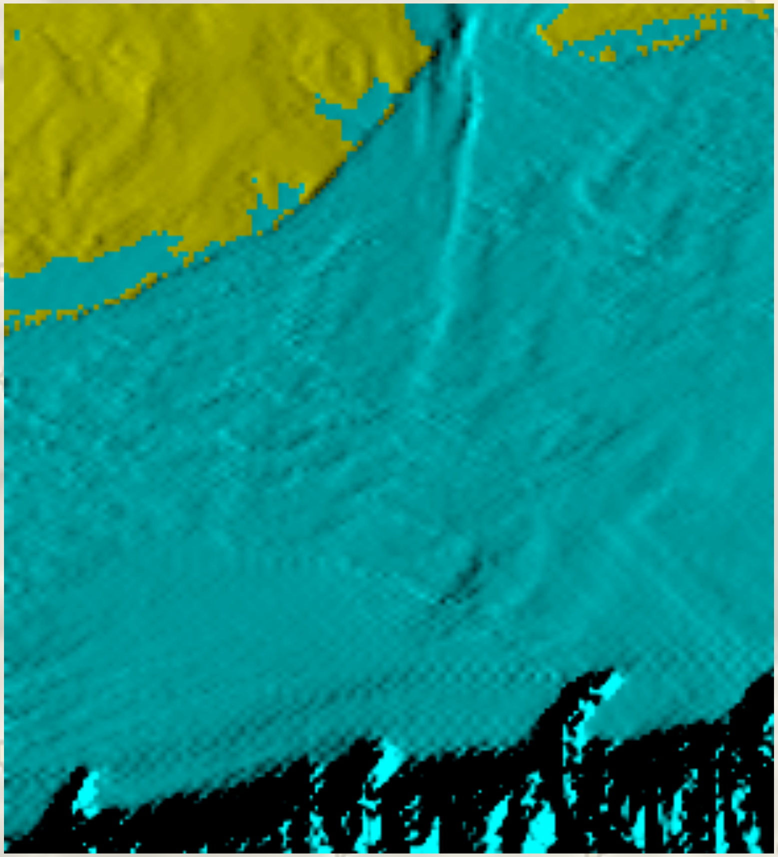
Frequent overwash *Infrequent overwash*



Modeling Shoreline Development



Without the coastal processes model the beach profile does not migrate. Barrier islands do not form



With the coastal processes model the beach profile follows the sea level. Barrier islands form during transgression

Sergio Fagherazzi

Fluvial Hypopycnal Plumes

after Syvitski et al.

- Steady 2D advection-diffusion equation:

$$\frac{\partial uI}{\partial x} + \frac{\partial vI}{\partial y} + \lambda I = \frac{\partial}{\partial y}\left(K \frac{\partial I}{\partial y} \right) + \frac{\partial}{\partial x}\left(K \frac{\partial I}{\partial x} \right)$$

Where: x, y are coordinate directions

u, v are velocities

K is turbulent sediment diffusivity

I is sediment inventory

λ is the first-order removal rate constant

Fluvial Hypopycnal Plumes

- Position of centerline: $$\frac{x}{b_0} = 1.53 + 0.90\left(\frac{u_0}{v_0}\right)\left(\frac{y}{b_0}\right)^{0.37}$$

- Concentration along and around centerline: $$C(x,y) = C_0 e^{-\lambda t}\sqrt{\frac{b_0}{\sqrt{\pi}C_1 x}e^{-\left(\frac{y}{\sqrt{2}C_1 x}\right)^2}}$$

where:

b_0=river mouth width

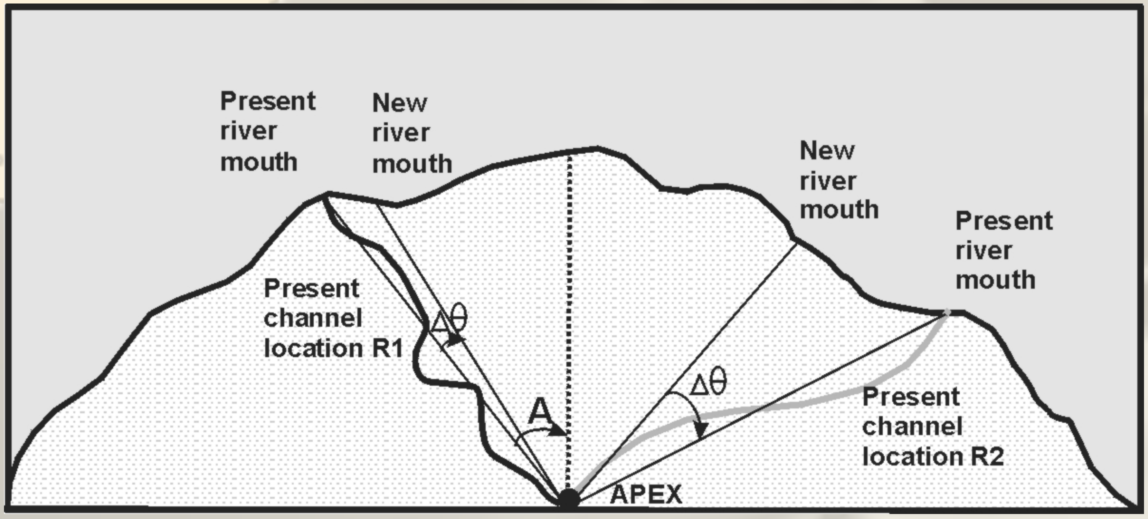
C_1=0.109 (empirical)

$$t(x,y) = \frac{u_0 + u_c(x) + 7u(x,y)}{9}$$

$$u_c(x) = u_0\sqrt{\frac{b_0}{\sqrt{\pi}C_1 x}}$$

$$u(x,y) = u_0\sqrt{\frac{b_0}{\sqrt{\pi}C_1 x}e^{-\left(\frac{y}{\sqrt{2}C_1 x}\right)^2}}$$

Avulsion mechanism: after Hutton & Syvitski



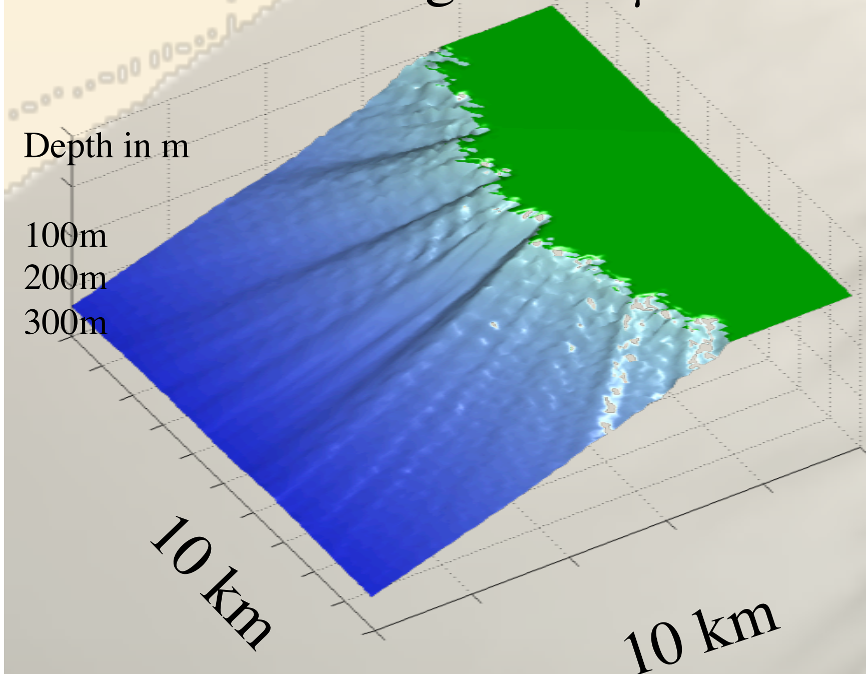
$$A_{t+\Delta t} = A_t + \Delta\theta$$

$$\Delta\theta = \mu X$$

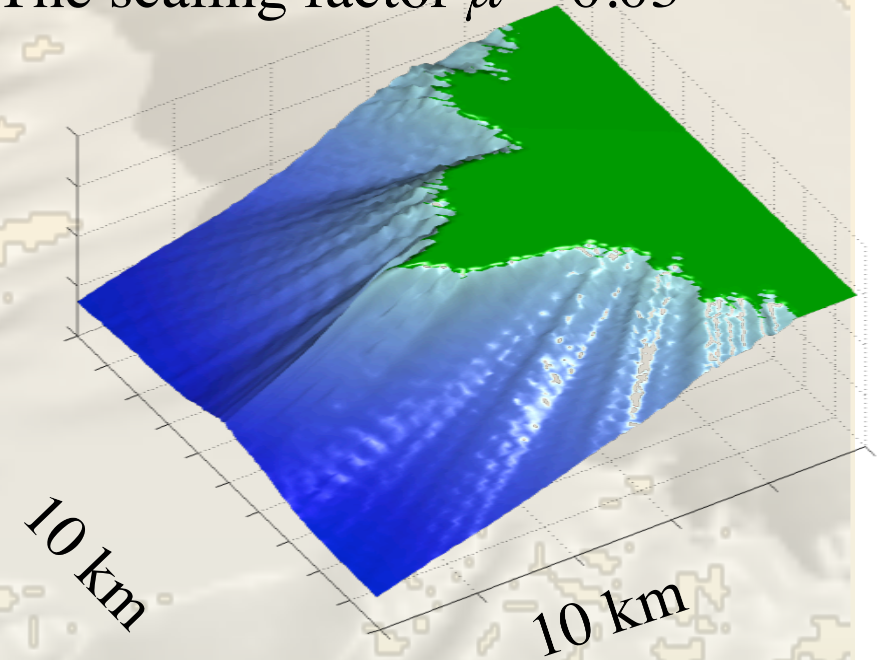
At specific time steps, $t + \Delta t$, the river mouth angle, A, changes by an amount drawn from a Gaussian distribution. The rate of switching is controlled by changing the scaling factor, μ of the Gaussian deviate, X.

Avulsions of main delta lobe

The scaling factor $\mu = 0.3$



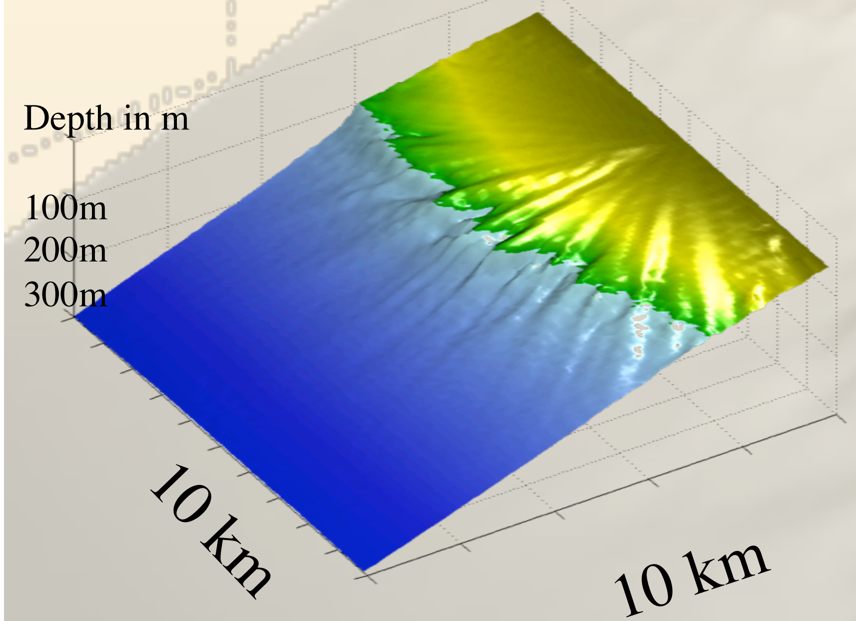
The scaling factor $\mu = 0.03$

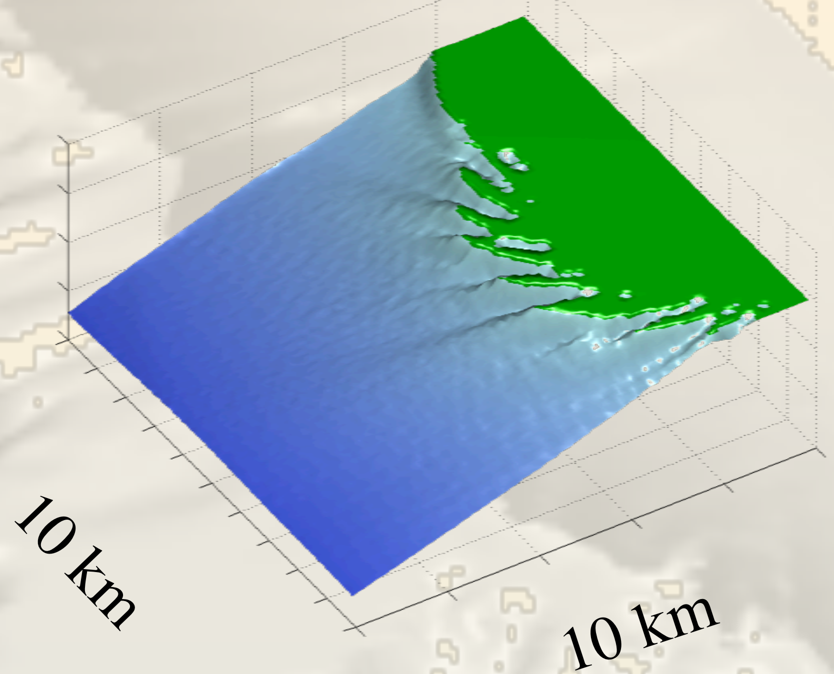


High avulsion frequency results in uniform progradation

Low avulsion frequency results in distinct lobe formation and locally enhanced progradation

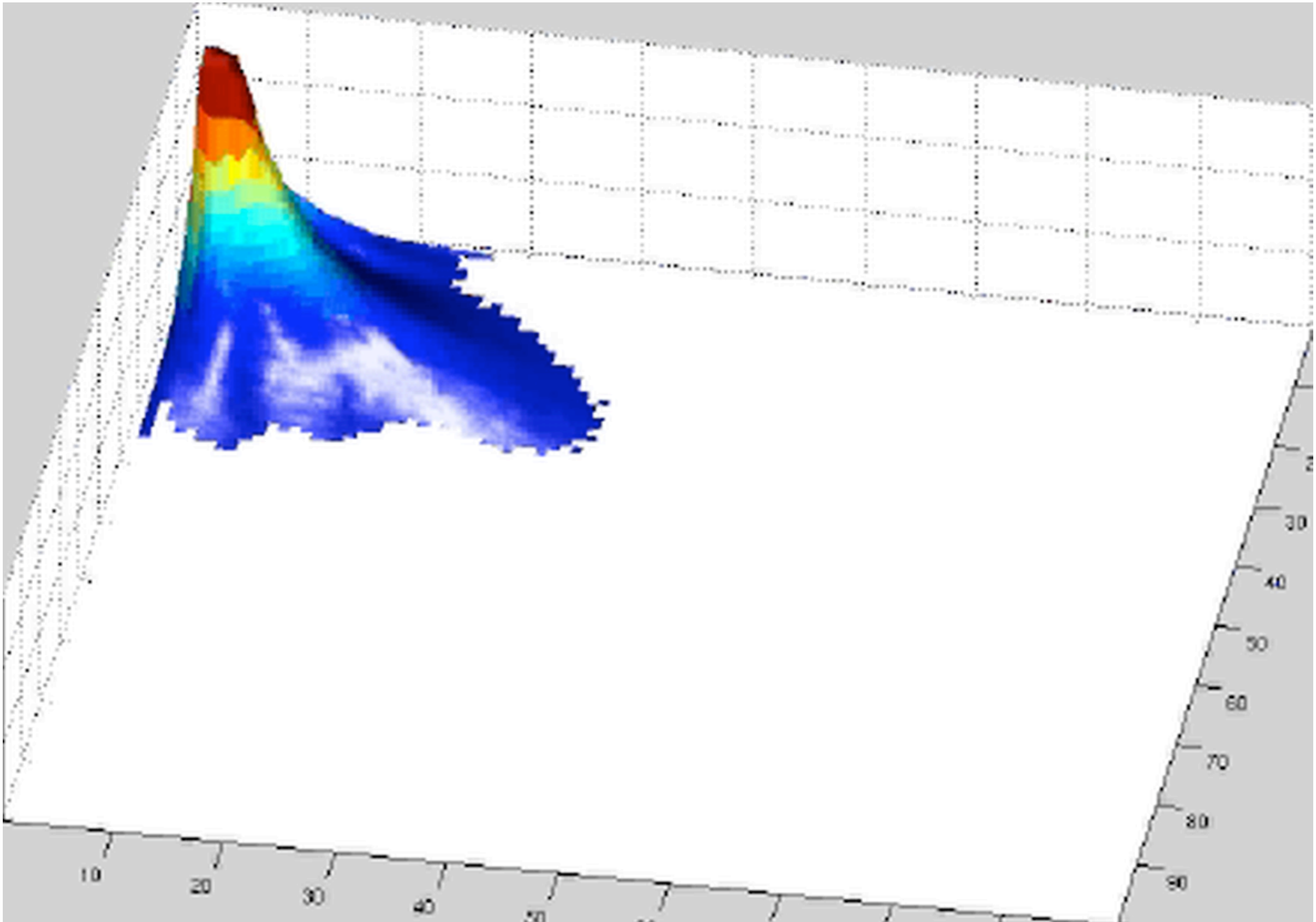
Plume forcing factors (SedFlux 3D)





Asymmetric progradation due to Coriolis Force

Curved plumes due to dominant current (0.5 m/s)


10
20
30
40
50
30
40
50
60
70
80
90

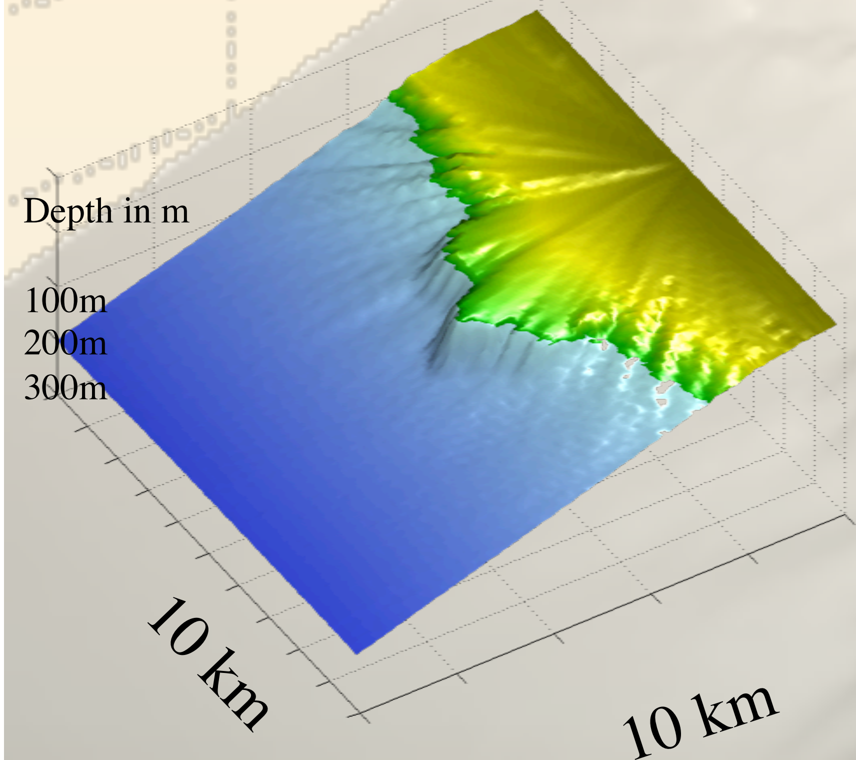
Fluvial erosion and deposition

Paola et al. (1992), using mass and momentum conservation

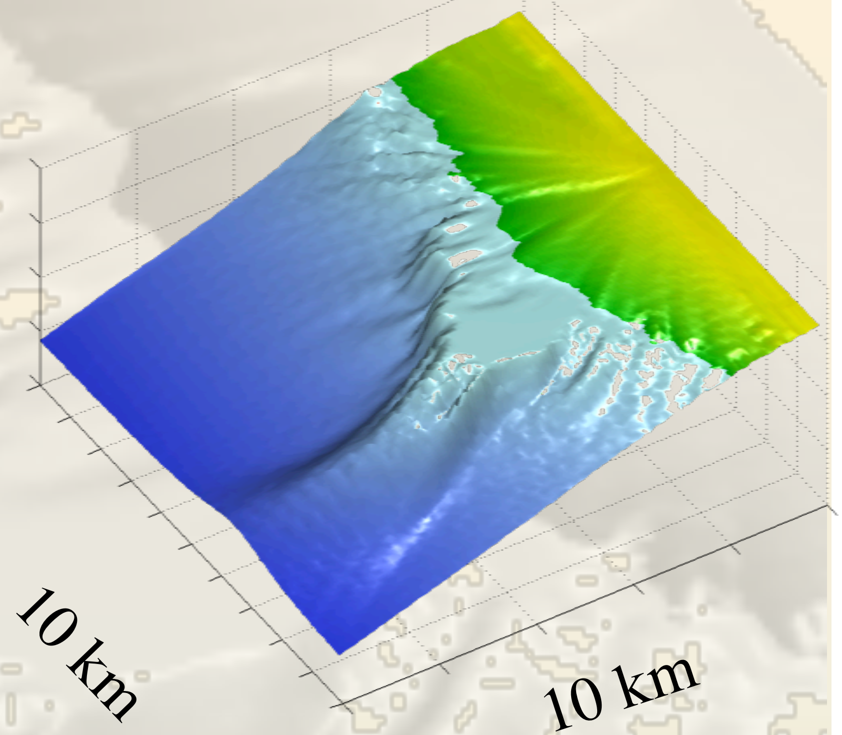
$$\frac{\delta\eta}{\delta t} = \nu \frac{\delta^2 \eta}{\delta x^2} \qquad \nu = \frac{-8\langle q \rangle A \sqrt{c_f}}{C_0 (s-1)}$$

- η is the height of the bed, t is time, x is the position along the long profile and ν is the diffusion coefficient.
- q is the discharge, cf is the drag coefficient, C_0 is the sediment concentration of the bed, s is sediment-specific gravity and A is a river-type dependent constant. A differs for meandering rivers (A=1) and braided rivers ($A = (\varepsilon / (1+ \varepsilon))$ 3/2). The value of ε relates the shear stress, τ, in the center of the channel to the critical shear stress, τ_c, needed for bank erosion ($\tau = (1+ \varepsilon)\, \tau_c$).

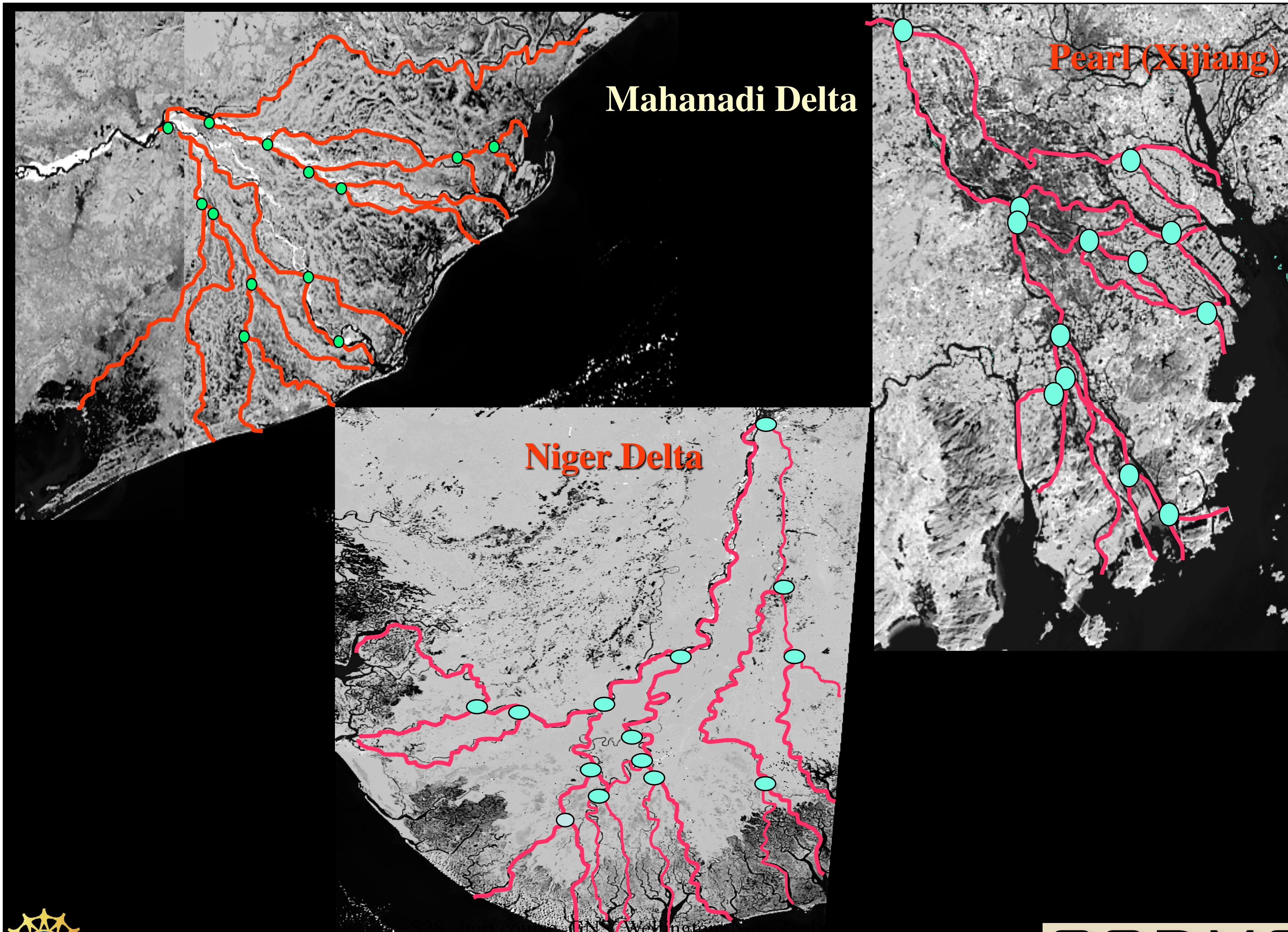
Fluvial erosion and deposition over full sea level cycle

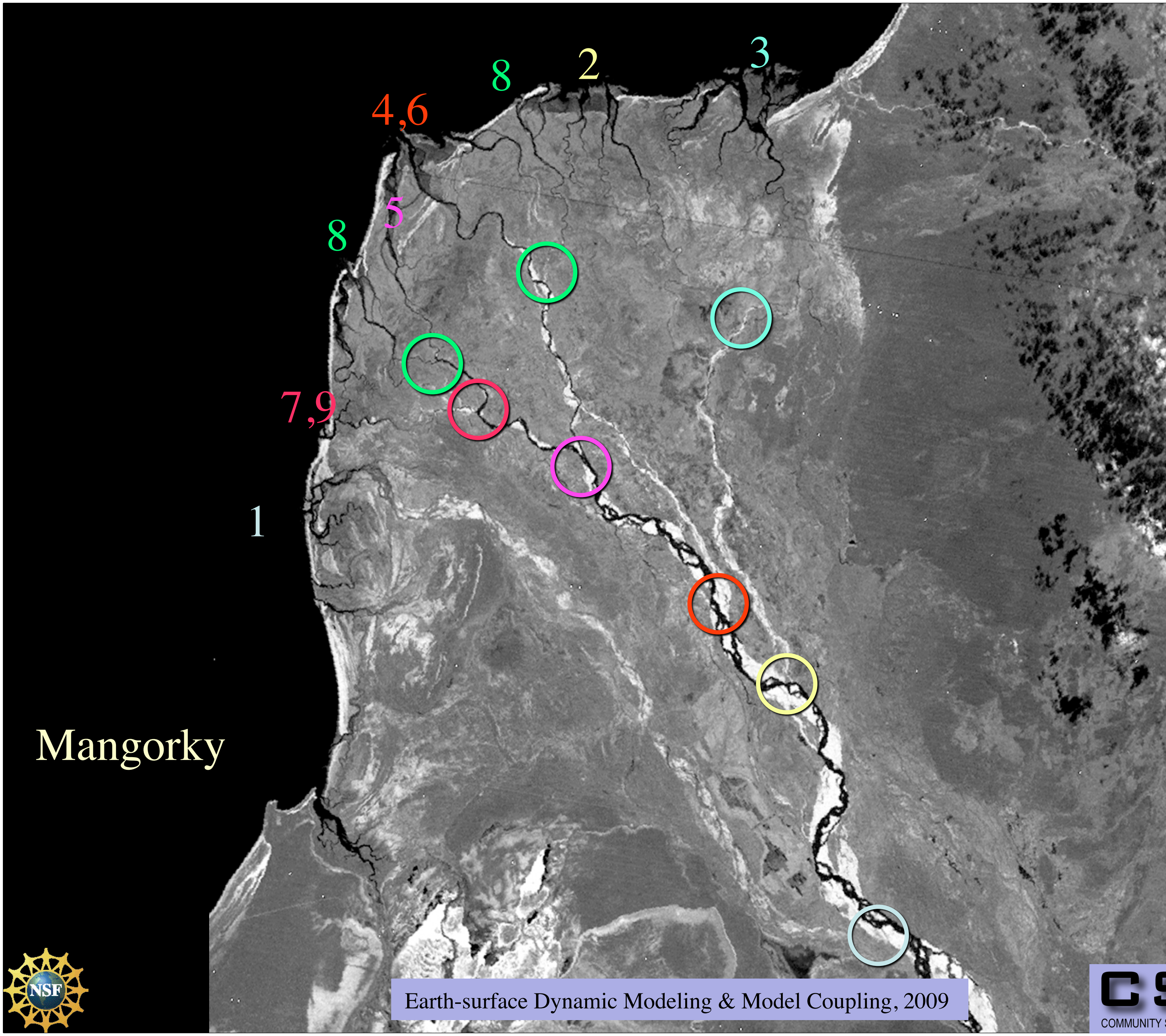


Lowstand conditions: several incised valleys developed



Highstand conditions: incised valleys are being backfilled, flooding surface develops


Mahanadi Delta
Pearl (Xijiang)
Niger Delta



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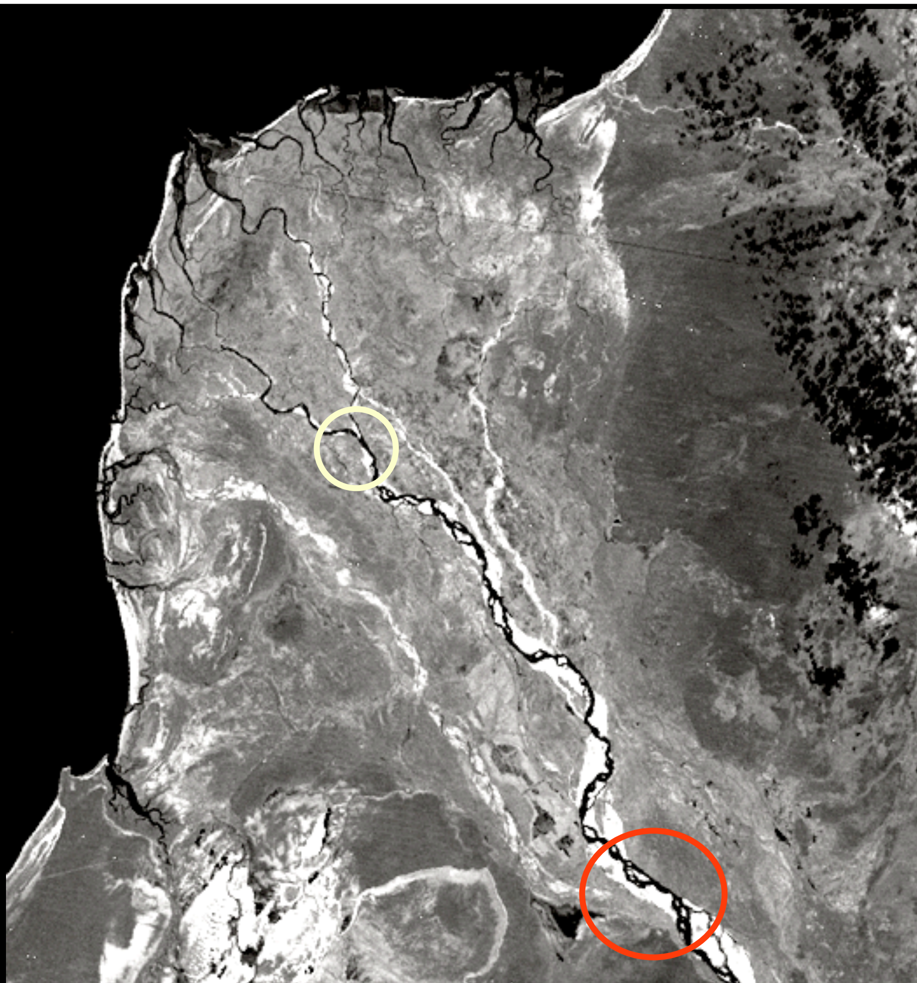
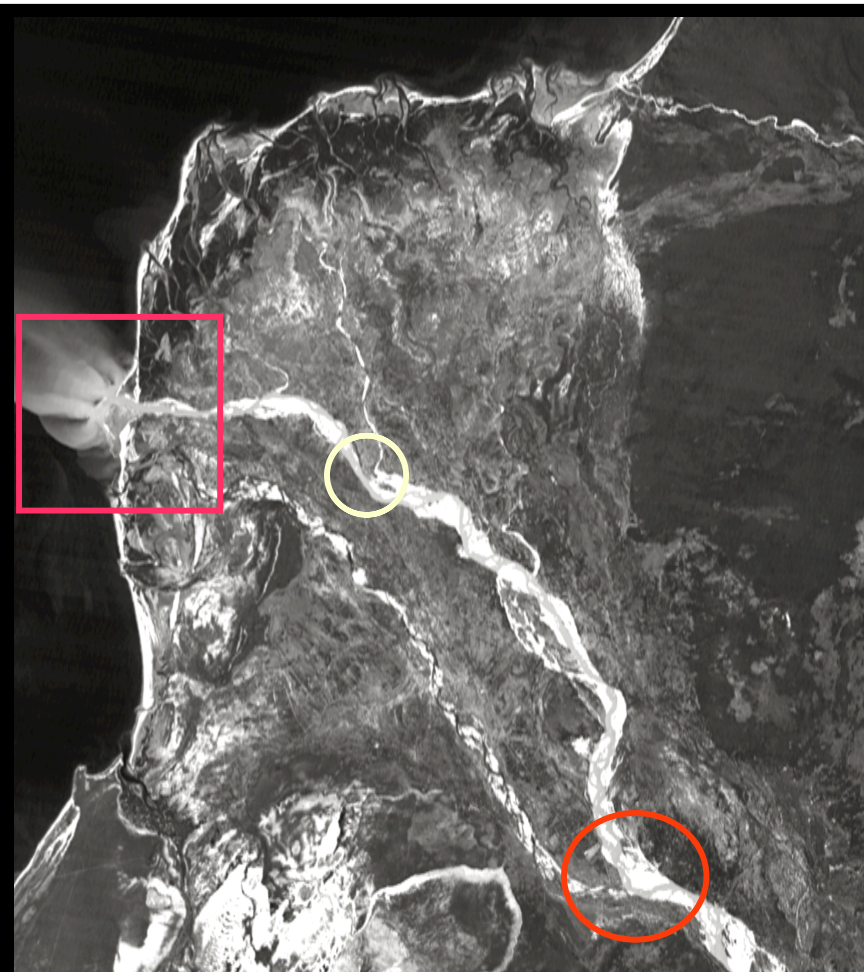
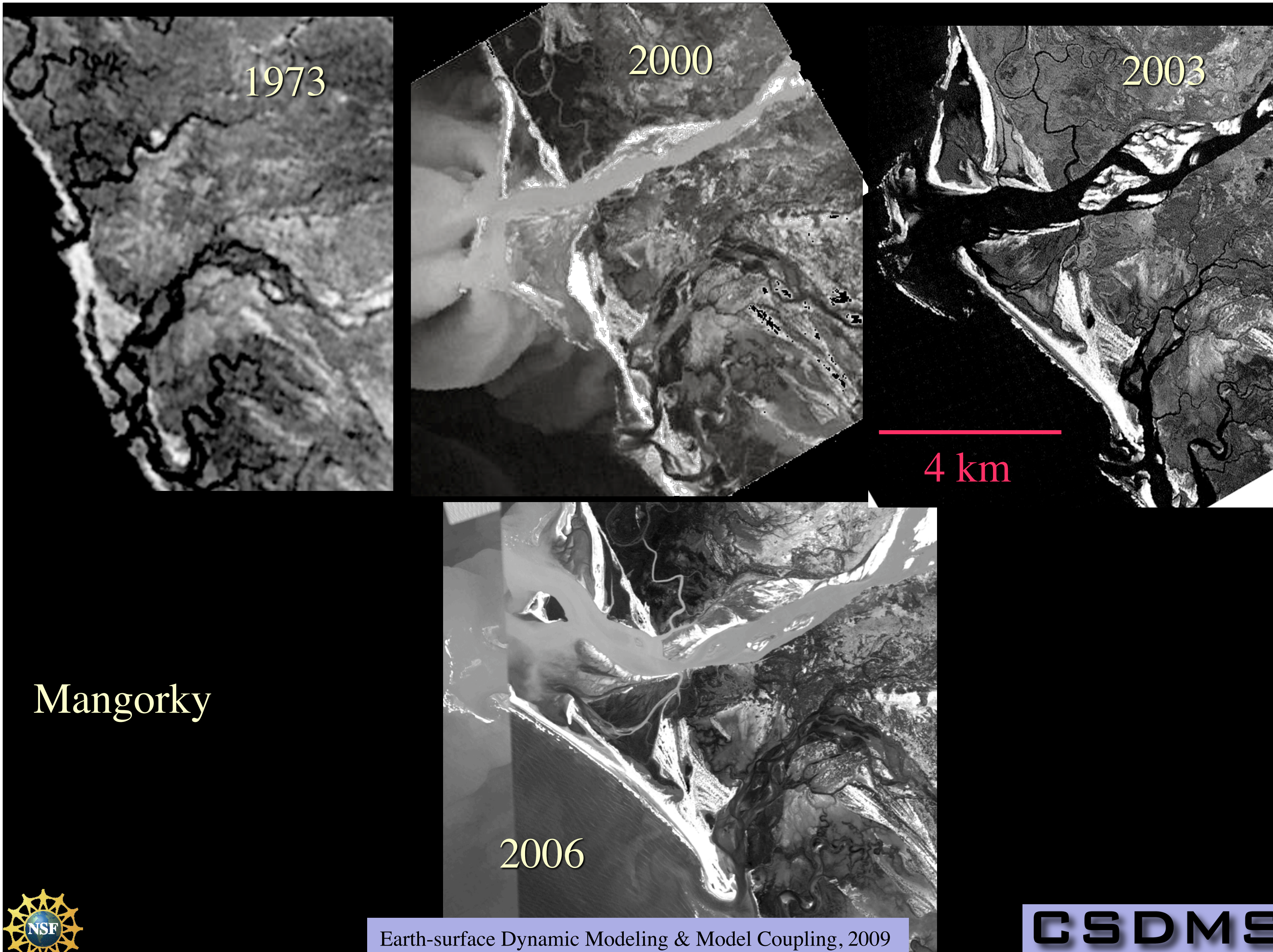


1973



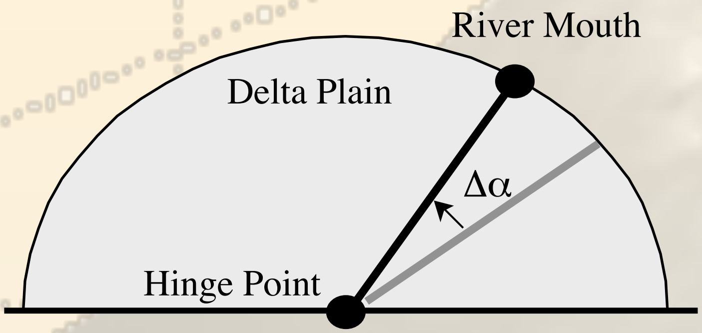
2003

Mangorky



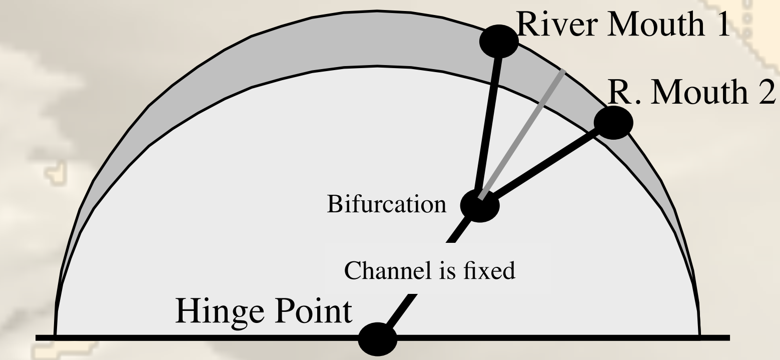
Mangorky

Channel Avulsion



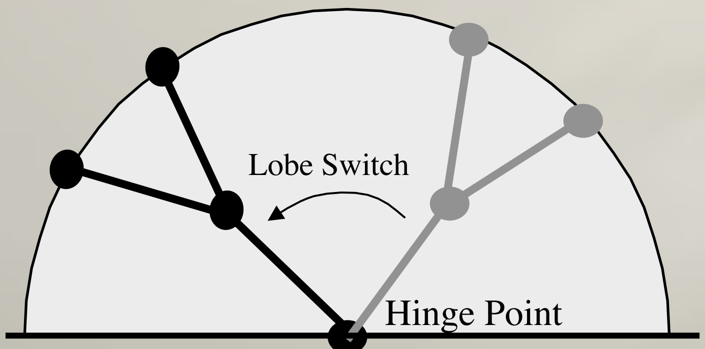
- Channel avulses about its hinge point.
- The avulsion angle (Δα) is drawn from a normal distribution.

Channel Bifurcation

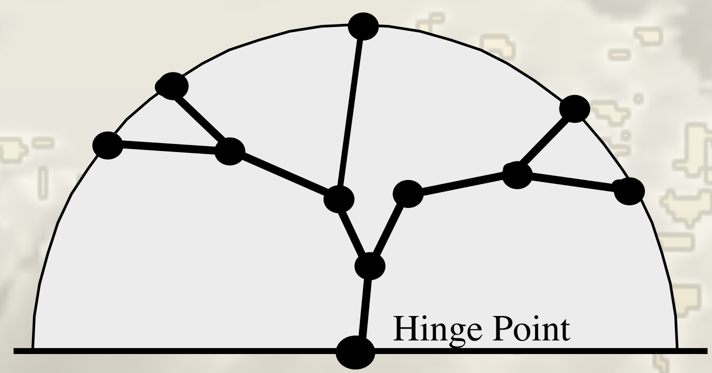


- If the ratio of delta area to number of channels exceeds a threshold, a channel bifurcates.
- The distributary channels avulse about the bifurcation point.
- The main channel is fixed for small Δα.

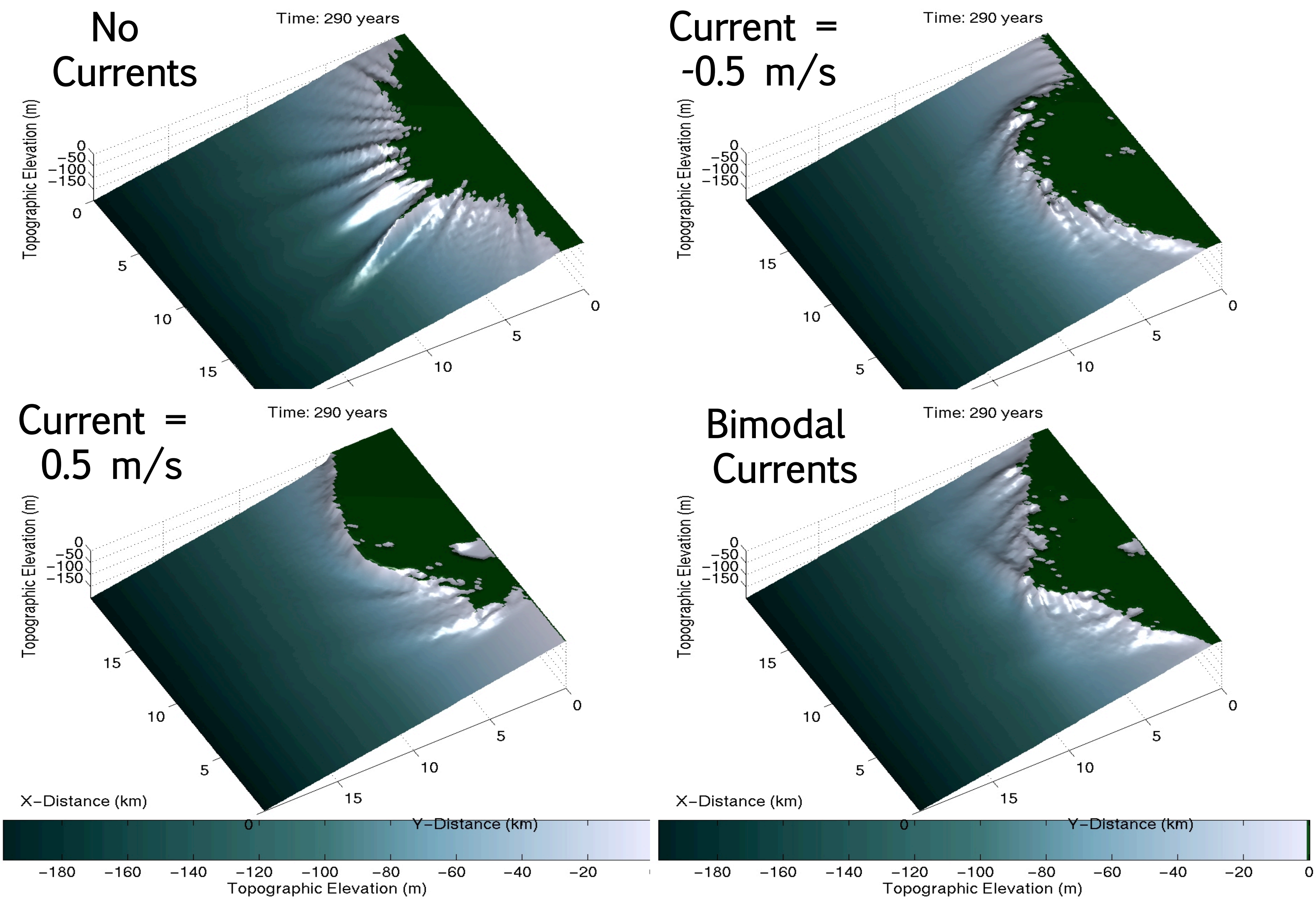
Lobe Switch

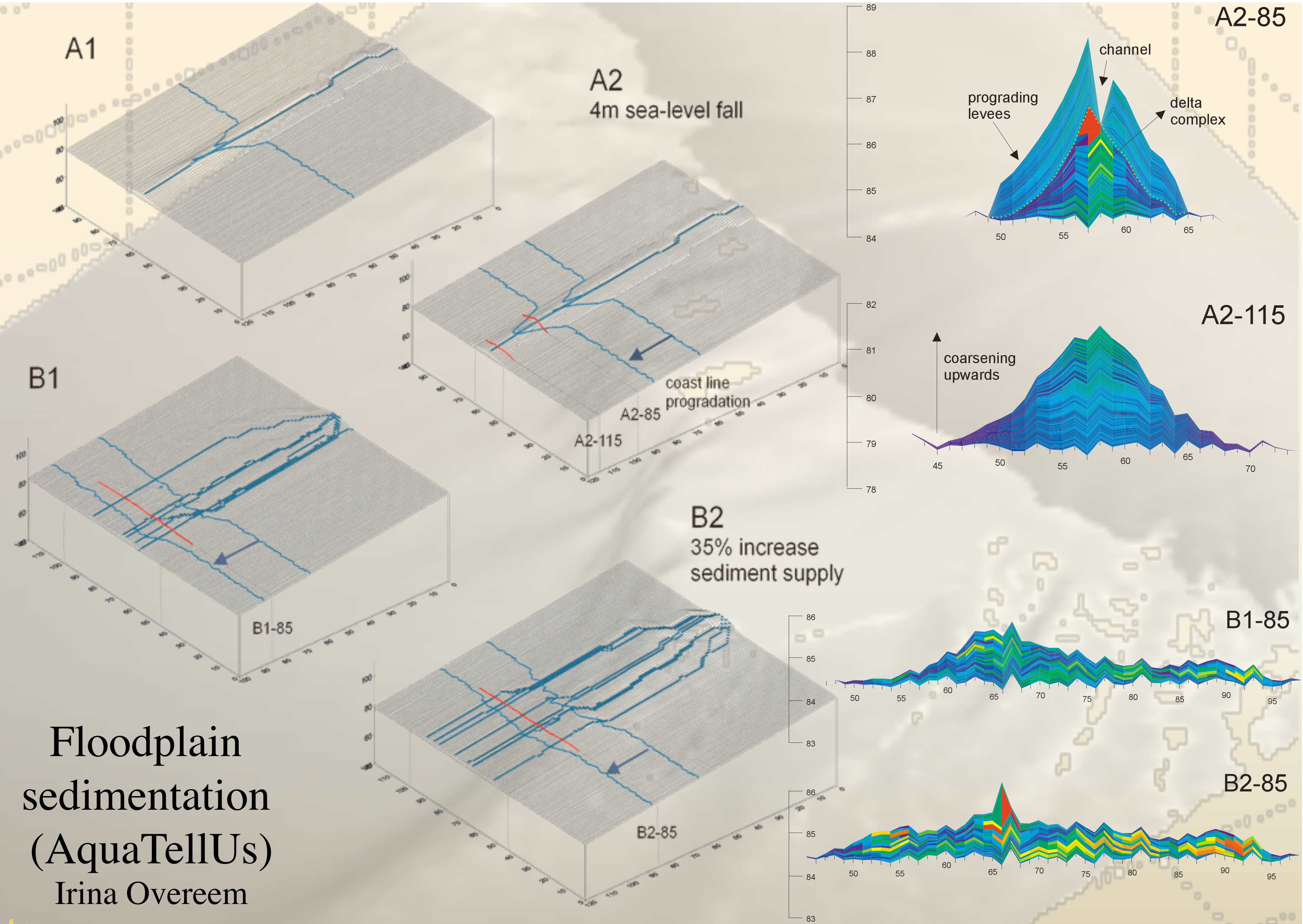


- If Δα exceeds a threshold, a lobe switch occurs
- The main channel is again fixed at its new location.
- Distributary channels avulse about their new bifurcation point.



- As the delta area grows more distributary channels are created.

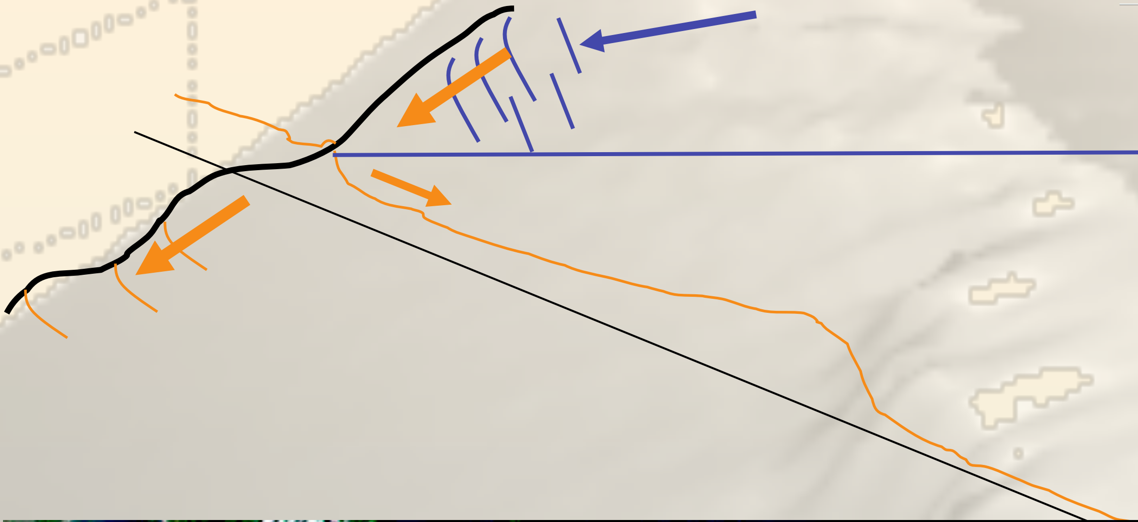

No Currents
Time: 290 years
Current = -0.5 m/s
Time: 290 years
Current = 0.5 m/s
Time: 290 years
Bimodal Currents
Time: 290 years
Topographic Elevation (m)
X-Distance (km)
Y-Distance (km)
Topographic Elevation (m)
−180 −160 −140 −120 −100 −80 −60 −40 −20 0

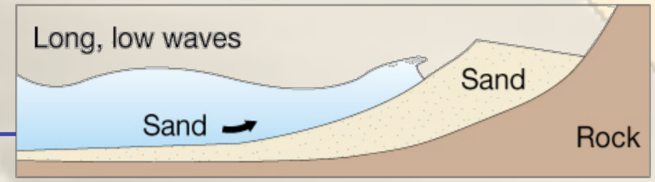


Floodplain sedimentation (AquaTellUs)

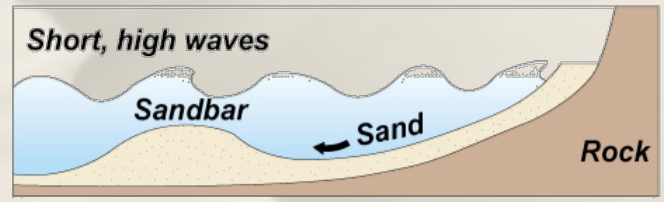
Irina Overeem

Along-Shore & Cross-Shore Sediment Transport





Summertime beach (fair weather)



Wintertime beach (storm)

Conservation of mass (Swenson et al., 2005)

$$\frac{\partial \eta}{\partial t} = \frac{16}{3\pi}\frac{\rho}{\rho_s - \rho}\frac{C_{fs}\varepsilon_{ss}}{g} I_s \left[\frac{\partial}{\partial x}\left(\frac{U^3}{w_s} v_0\right) + \frac{\partial}{\partial x}\left(\frac{U^5}{5w_s^2}\frac{\partial \eta}{\partial x}\right)\right]$$

Resuspension Criteria (Komar, 1998)

$$\frac{\rho U^2}{(\rho_s - \rho) g D} = \begin{cases} .21\left(\frac{d_0}{D}\right)^{1/2} & \text{for } D \le .5\text{mm} \\ .46\pi\left(\frac{d_0}{D}\right)^{1/4} & \text{for } D > .5\text{mm} \end{cases}$$

Conservation of wave action (Mei, 1989)

$$\left(\frac{E}{\sigma}\right)_t + \left[U + C_g \cos(\theta)\frac{E}{\sigma}\right]_x + \left[V + C_g \sin(\theta)\frac{E}{\sigma}\right]_y = -\frac{\varepsilon_b}{\sigma}$$

$$\varepsilon_b = \frac{3\sqrt{\pi}}{16}\rho g f_p d^2 (B\gamma w)^3 \left\{1 + \tanh\left[8(w-1)\right]\right\}\left[1 - \left(1 + w^2\right)^{-5/2}\right]$$

$$w = \frac{H_{rms}}{\gamma d}$$

$$U(\eta) = \frac{\gamma_b}{2}\sqrt{g\eta_b}\left(\frac{\eta}{\eta_b}\right)^{-3/4}$$ Shoaling Waves

$$U = \frac{\pi d_0}{T} = \frac{\pi H}{T \sinh(2\pi h/L)}$$ Deeper Waves

$$c \equiv \frac{\omega}{\kappa} \qquad \frac{c}{c_\infty} = \tanh(\kappa h)$$ Phase velocity

$$2n \equiv 1 + \frac{2\kappa h}{\sinh(2\kappa h)}$$ Definition of n

$$\frac{H}{H_\infty} = \sqrt{\frac{1}{2n}\frac{c_\infty}{c}} = \sqrt{\frac{\cosh(2\kappa h) + 1}{\sinh(2\kappa h) + 2\kappa h}}$$ Wave height

$$H = .00195\left(\frac{T}{.245}\right)^{2.5}$$ Fully-developed sea waves

$$\omega^2 = g\kappa \tanh(\kappa h) \Rightarrow \kappa = \frac{\omega^2}{g}\coth(\kappa h)$$ Dispersion

$$U_{\text{bed}} = \frac{\omega H}{2\sinh(\kappa h)}$$ Near-bottom orbital velocity

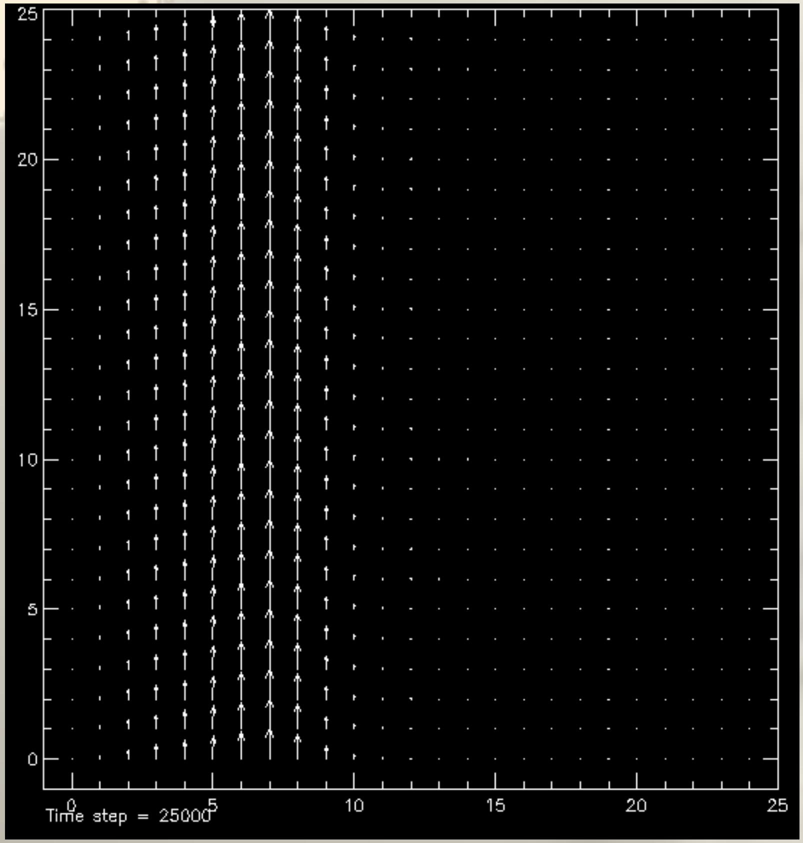
$$\left.\begin{matrix} U \sim h^{.5} \\ q \sim U^5 \end{matrix}\right\} \Rightarrow q \sim h^{2.5} \qquad f(q) = f(h^{2.5})$$ Scaling

$$z = a(y - y_{shore})^{2/3}$$ Breaker zone profile

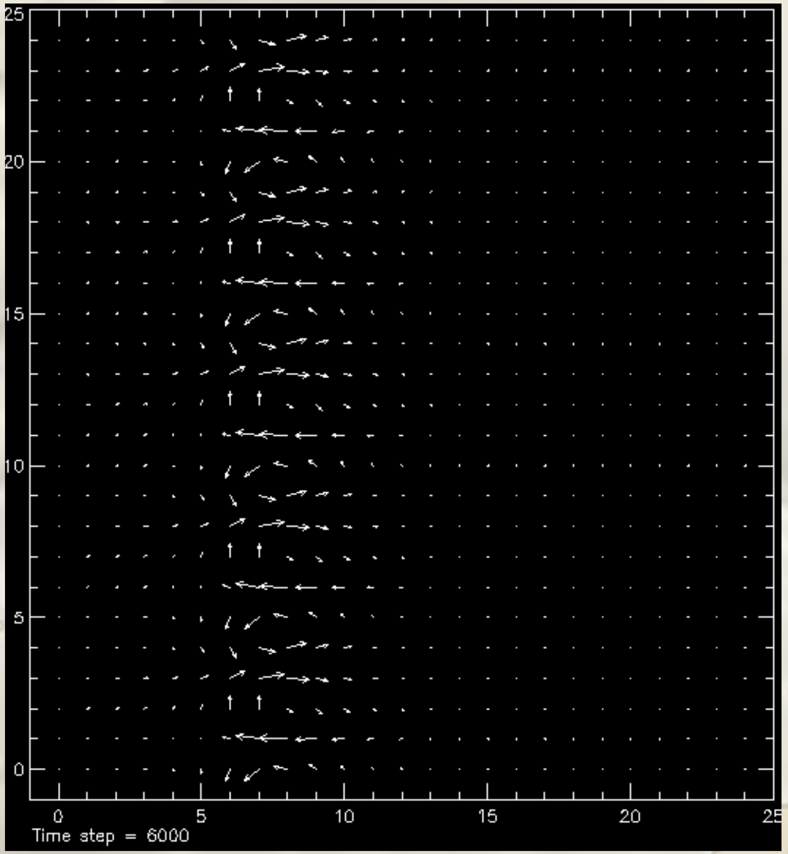
$$[g\,k\,\tanh(k\,d)]^{1/2} + k\,[U\cos(\theta) + V\sin(\theta)] = \frac{2\pi}{T}$$ Wave-current interaction

Wave-Induced Currents

Longshore velocity
vs. distance offshore m

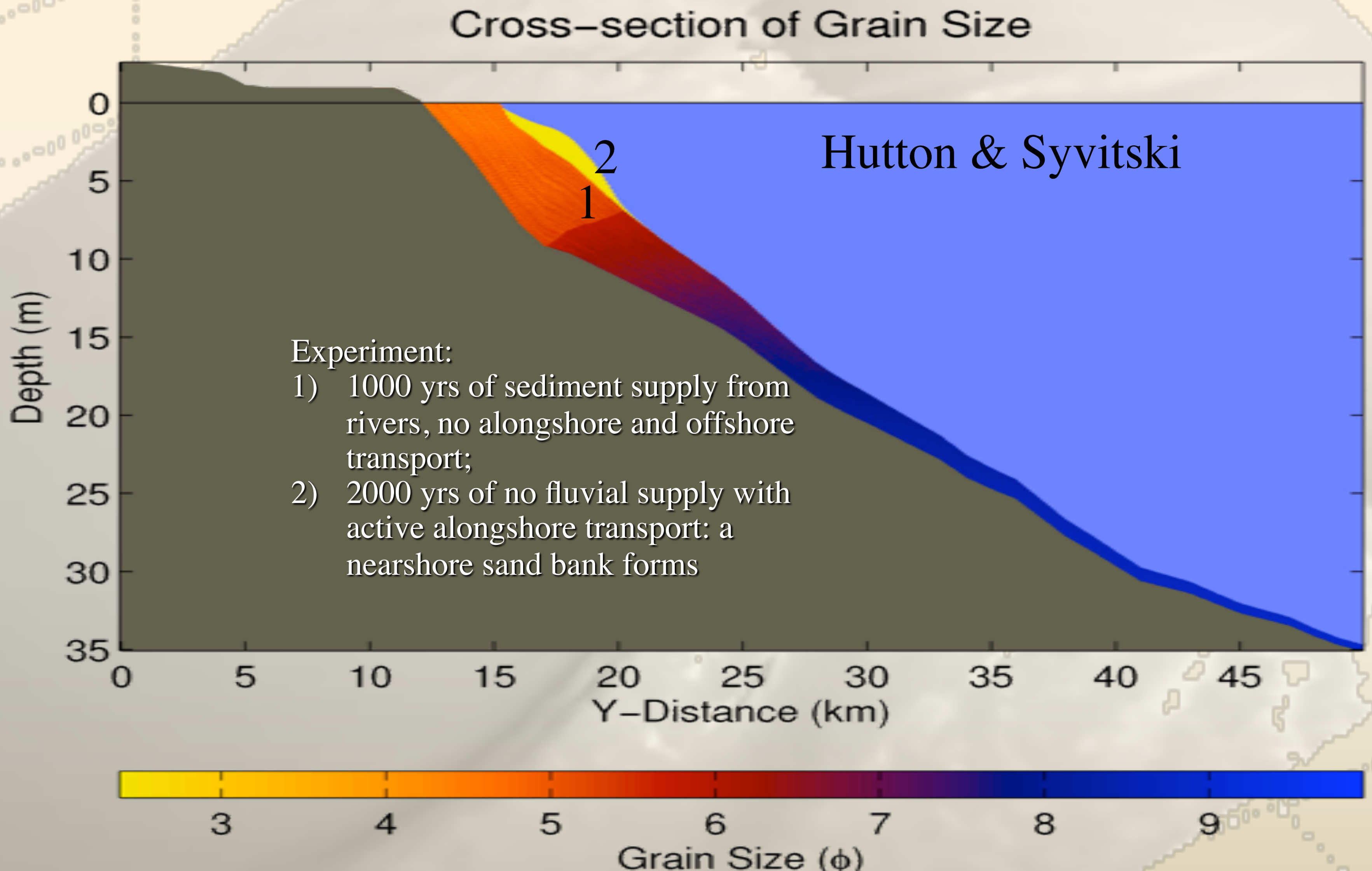


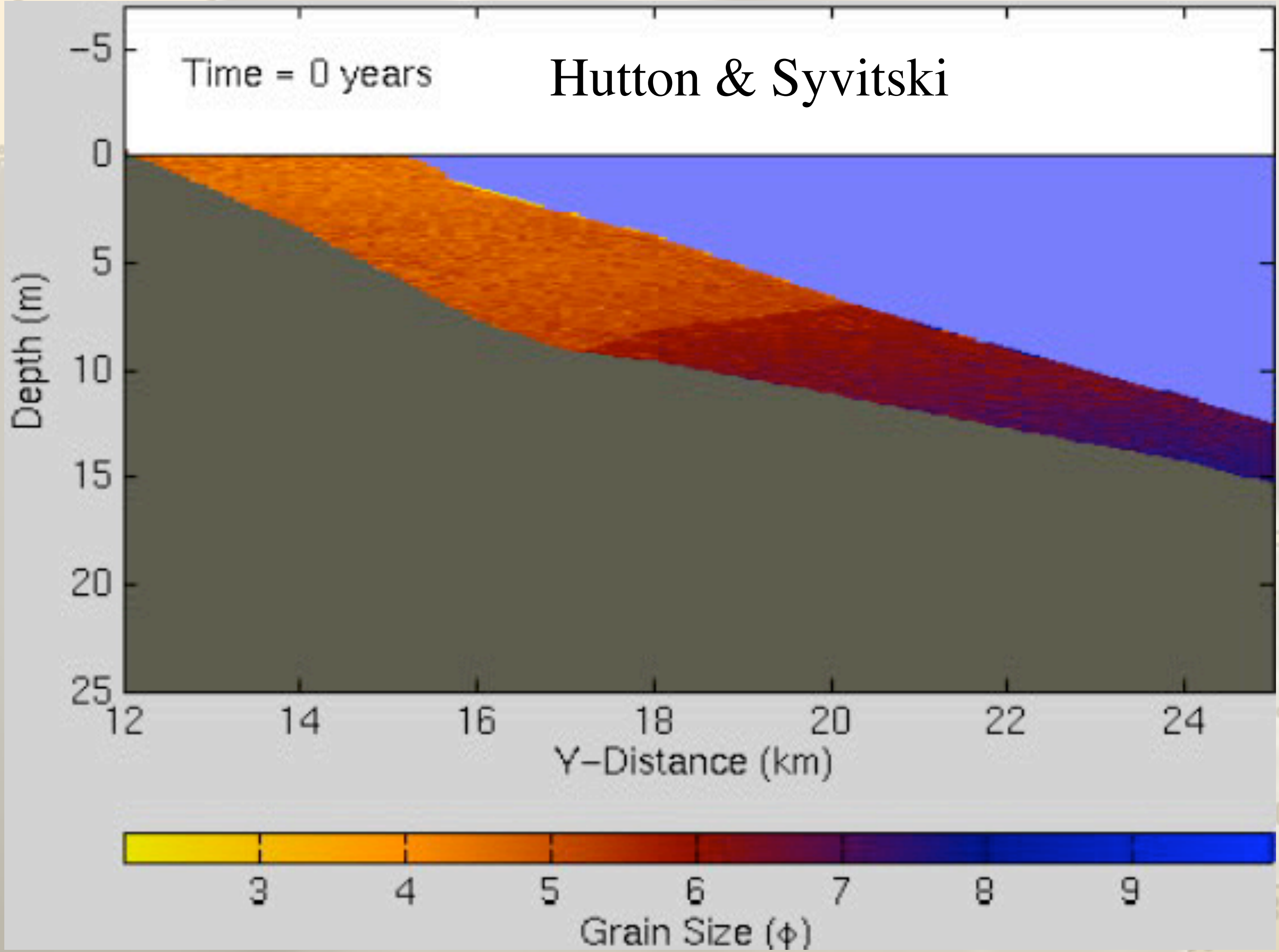
Cross-shore currents
vs. distance offshore m



Courtesy S. Peckham, CSDMS

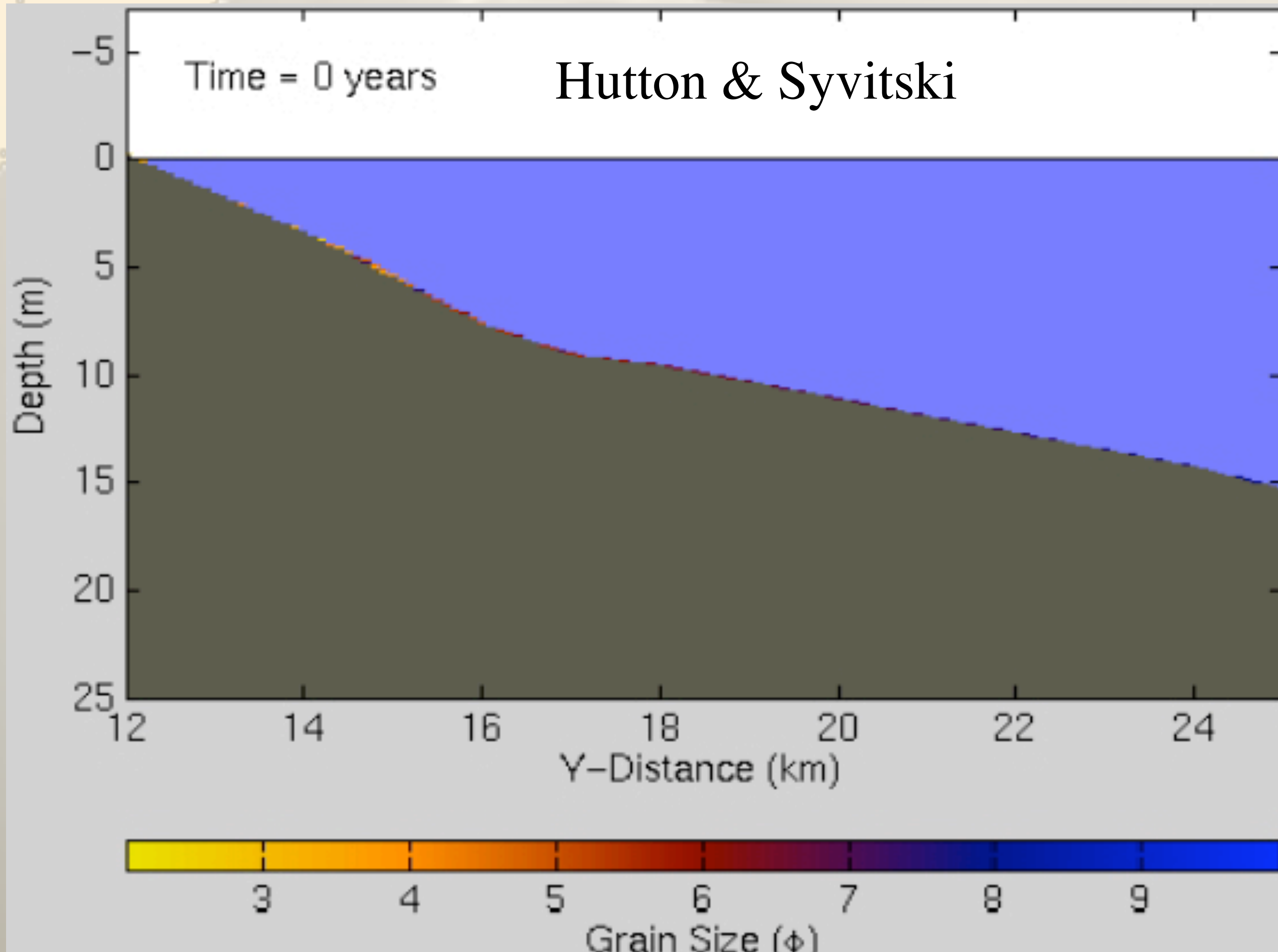
Alongshore Transport of Sand (0.2 mm)





Time = 0 years
Hutton & Syvitski
-5
0
5
10
15
20
25
Depth (m)
12
14
16
18
20
22
24
Y-Distance (km)
3
4
5
6
7
8
9
Grain Size (ϕ)

Fluvial Sediment Supply and Alongshore Transport



Coastal Morphodynamics Summary

Relative Sea Level: complicated, involving many processes, often with time lags, including eustasy, isostasy, compaction, humans, tectonics, subsidence

Drainage Basins & Sea Level: time and space variability, influence by initial topography, razor-blade effect during rising SL

Barrier Islands & Sea Level: beach profiles, overwash, SL rise speed,

Deltas & Channel Switching: discharge plumes (shelf currents, Coriolis), channel avulsions, aggradation, sea level, bifurcation, steepest descent

Alongshore & Cross-Shore Transport Modeling: alongshore & cross-shore velocity, wave dynamics, beach sands.