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Damping of turbulence by suspended sediment: fundamental ramifications for sediment dynamics

Carl T. Friedrichs
Virginia Institute of Marine Science

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Damping of Turbulence by Suspended Sediment: Fundamental Ramifications for Sediment Dynamics

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<u>Outline of Presentation</u>:

- **Richardson number influence on coastal/estuarine mixing**
- **Derivation of stratified "overlap" layer structure**
- **Under-saturated (weakly stratified) sediment suspensions**
- **Critically saturated (Ri_{cr}-controlled) sediment suspensions**
- **Hindered settling, over-saturation, and collapse of turbulence**

Presented at University of Delaware, 9/27/11

10

1 meter

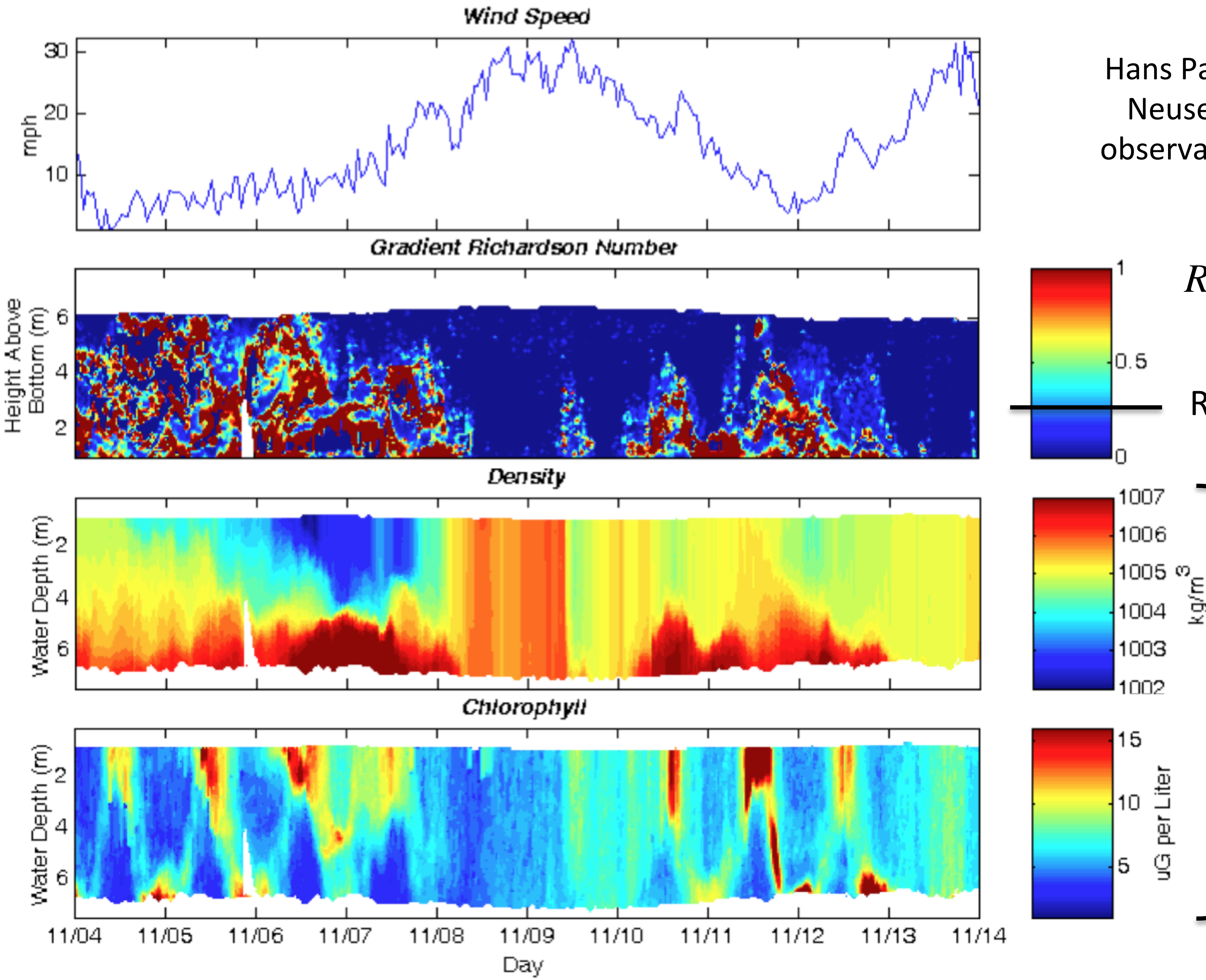
30

2 hours

100 mg/l Time-series of suspended sediment in York River estuary (Friedrichs et al. 2000)

Gradient Richardson Number (Ri) = $\frac{\text{density stratification}}{\text{velocity shear}}$

Shear instabilities occur for $Ri < Ri_{cr}$
" " suppressed for $Ri > Ri_{cr}$



Hans Paerl et al. (2004)
Neuse River Estuary
observations from 2003

$$Ri = \frac{g\ (\partial\rho/\partial z)}{\rho\ (\partial u/\partial z)^2}$$

$Ri_{cr} \approx 1/4$

Density & Chl stratified when $Ri > 1/4$, mixed when $Ri < 1/4$

OUTLINE: **1) Ri # importance**; 2) Overlap layer; 3) Under-saturation; 4) Critical Saturation; 5) Over-saturation

When strong currents are present, mud remains turbulent and in suspension at a concentration that gives $Ri \approx Ri_{cr} \approx 1/4$:

Gradient Richardson Number (Ri) = $\dfrac{\text{density stratification}}{\text{velocity shear}}$

Shear instabilities occur for $Ri < Ri_{cr}$
" " suppressed for $Ri > Ri_{cr}$

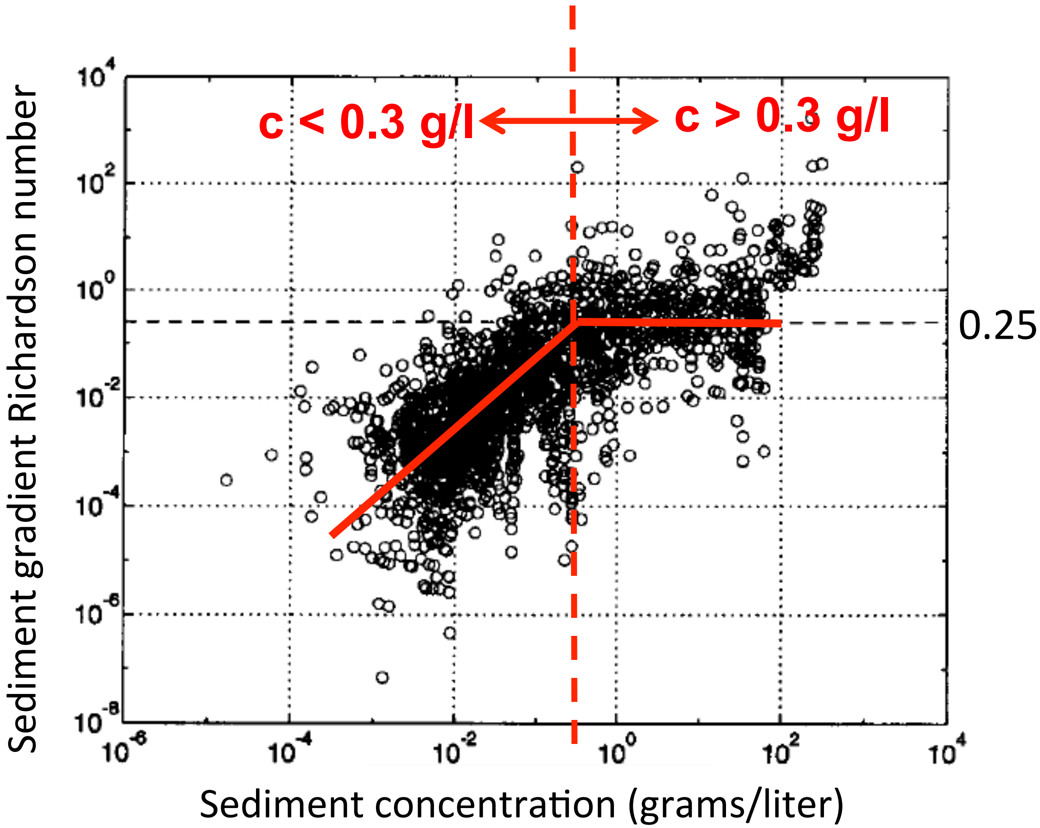


Figure 5. Sediment-based gradient Richardson number as a function of sediment concentration based on measurements throughout the entire water column in all of the profiles summarized in Table 1. The dashed curve corresponds to a gradient Richardson number of 1/4.

$$Ri = \frac{\text{Stratification}}{\text{Shear}}$$

$$Ri = \frac{-\,g\;s\;\partial c/\partial z}{\rho_s\,(\partial u/\partial z)^2}$$

g = accel. of gravity
$s = (\rho_s - \rho)/\rho$
c = sediment mass conc.
ρ_s = sediment density

For c > ~ 300 mg/liter

$Ri \approx Ri_{cr} \approx O(1/4)$

Amazon Shelf (Trowbridge & Kineke, 1994)

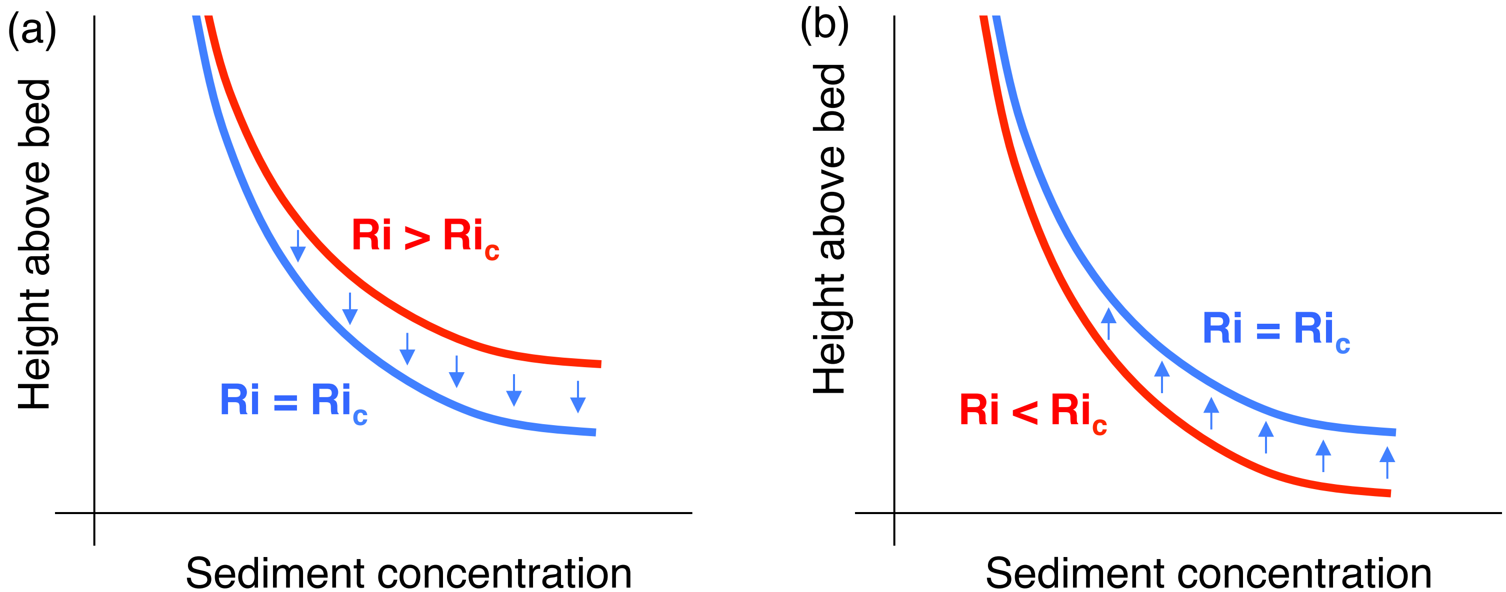
OUTLINE: **1) Ri # importance**; 2) Overlap layer; 3) Under-saturation; 4) Critical Saturation; 5) Over-saturation

Are there simple, physically-based relations to predict c and du/dz related to Ri?

Large supply of easily suspended sediment creates negative feedback:

Gradient Richardson Number (Ri) = $\frac{\text{density stratification}}{\text{velocity shear}}$

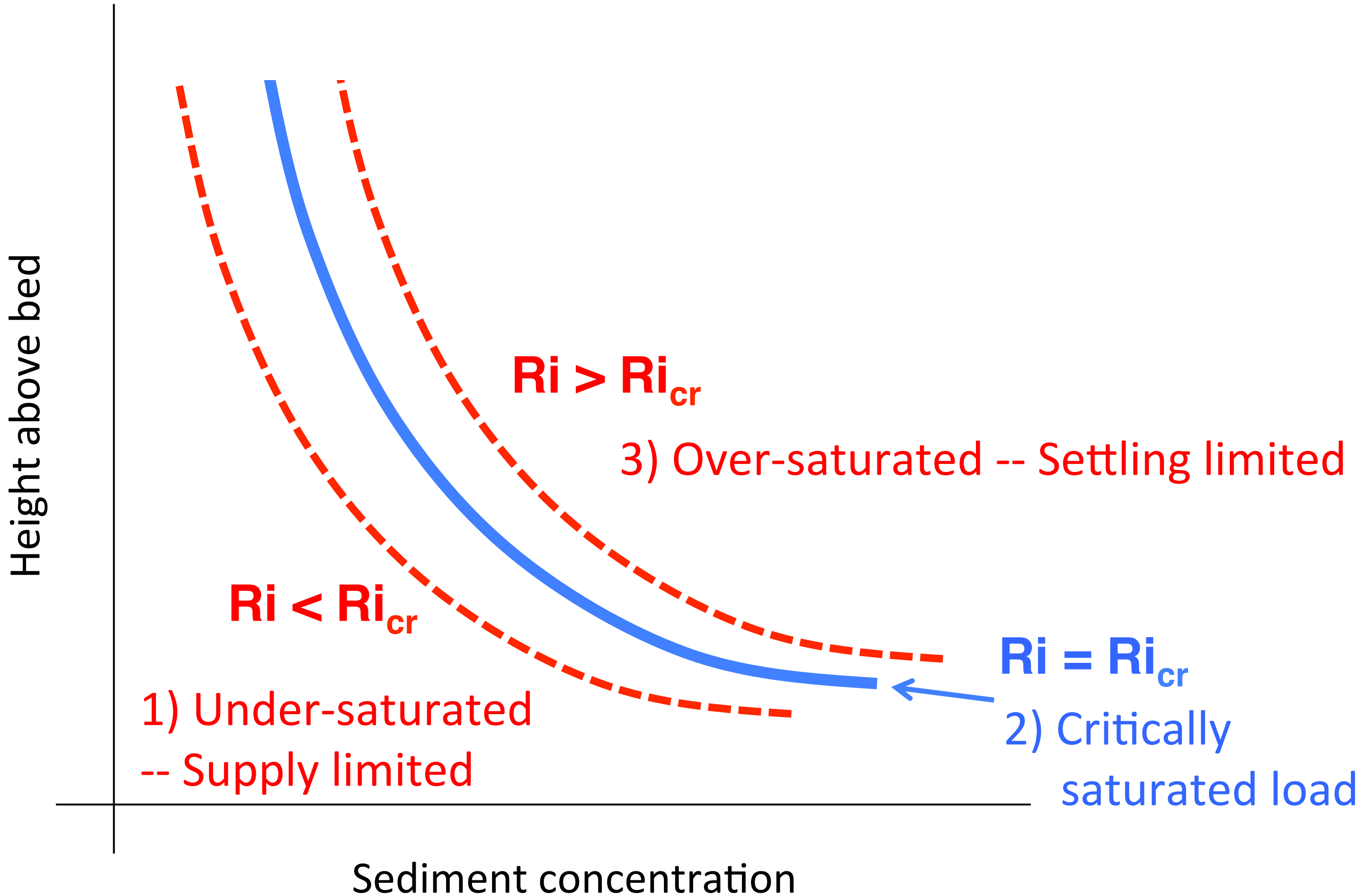
Shear instabilities occur for $Ri < Ri_{cr}$
" " suppressed for $Ri > Ri_{cr}$



(a) If excess sediment enters bottom boundary layer or bottom stress decreases, **Ri ↑ beyond Ri_c**, critically damping turbulence. Sediment settles out of boundary layer. Stratification is reduced and **Ri returns to Ri_c**.

(b) If excess sediment settles out of boundary layer or bottom stress increases, **Ri ↓ below Ri_c** and turbulence intensifies. Sediment re-enters base of boundary layer. Stratification is increased in lower boundary layer and **Ri returns to Ri_c**.

Consider Three Basic Types of Suspensions



OUTLINE: **<u>1) Ri # importance</u>**; 2) Overlap layer; 3) Under-saturation; 4) Critical Saturation; 5) Over-saturation

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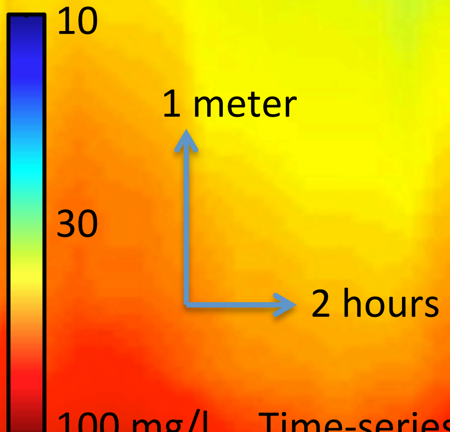
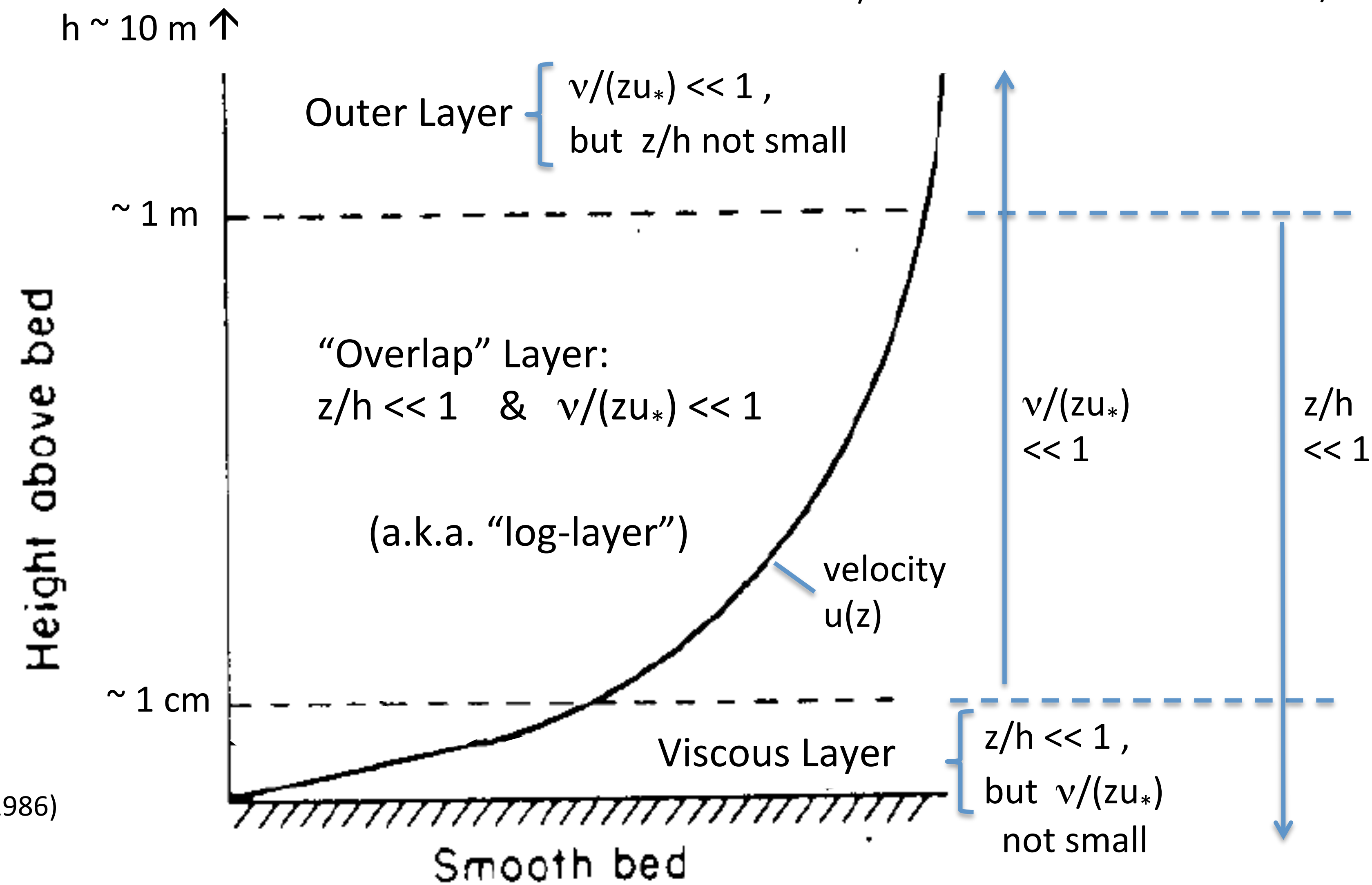


Time-series of suspended sediment in York River estuary (Friedrichs et al. 2000)

Dimensionless analysis of bottom boundary layer in the absence of stratification:

Variables
du/dz, z, h, ν, u_* → $$\frac{z}{u_*}\frac{du}{dz} = f\left(\frac{\nu}{zu_*}, \frac{z}{h}\right)$$

h = thickness of bottom boundary layer, ν = kinematic viscosity, ~ 10^{-6} m^2/s, $u_* = (\tau_b/\rho)^{1/2}$ = shear velocity ~ 1 cm/s



(Dyer, 1986)

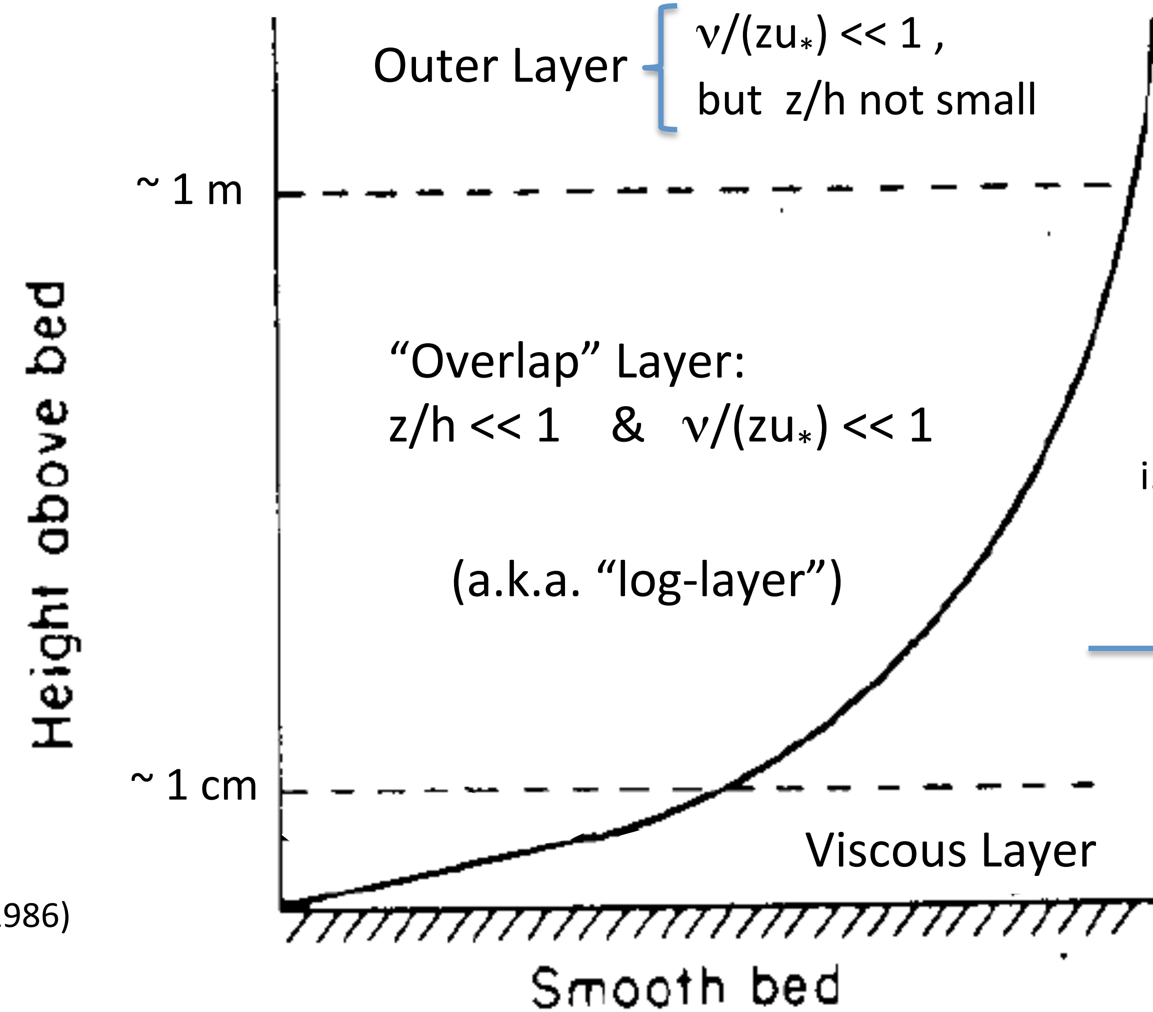
OUTLINE: 1) Ri # importance; **2) Overlap layer**; 3) Under-saturation; 4) Critical Saturation; 5) Over-saturation

Dimensionless analysis of bottom boundary layer in the absence of stratification:

Variables
du/dz, z, h, ν, u_* → $$\frac{z}{u_*}\frac{du}{dz} = f\left(\frac{\nu}{z u_*}, \frac{z}{h}\right)$$

h = thickness of bottom boundary layer, ν = kinematic viscosity, $u_* = (\tau_b/\rho)^{1/2}$ = shear velocity
~ 10^{-6} m²/s ~ 1 cm/s

h ~ 10 m ↑



(Dyer, 1986)

~ 1 m

Outer Layer – $\nu/(zu_*) << 1$, but z/h not small

$$\frac{z}{u_*}\frac{du}{dz} = f\left(\frac{z}{h}\right)$$

"Overlap" Layer:
z/h << 1 & $\nu/(zu_*) << 1$

(a.k.a. "log-layer")

$$\frac{z}{u_*}\frac{du}{dz} = \text{Const.} = 1/\kappa$$

i.e., $$\frac{\kappa z}{u_*}\frac{du}{dz} = 1 \qquad \kappa \approx 0.41$$

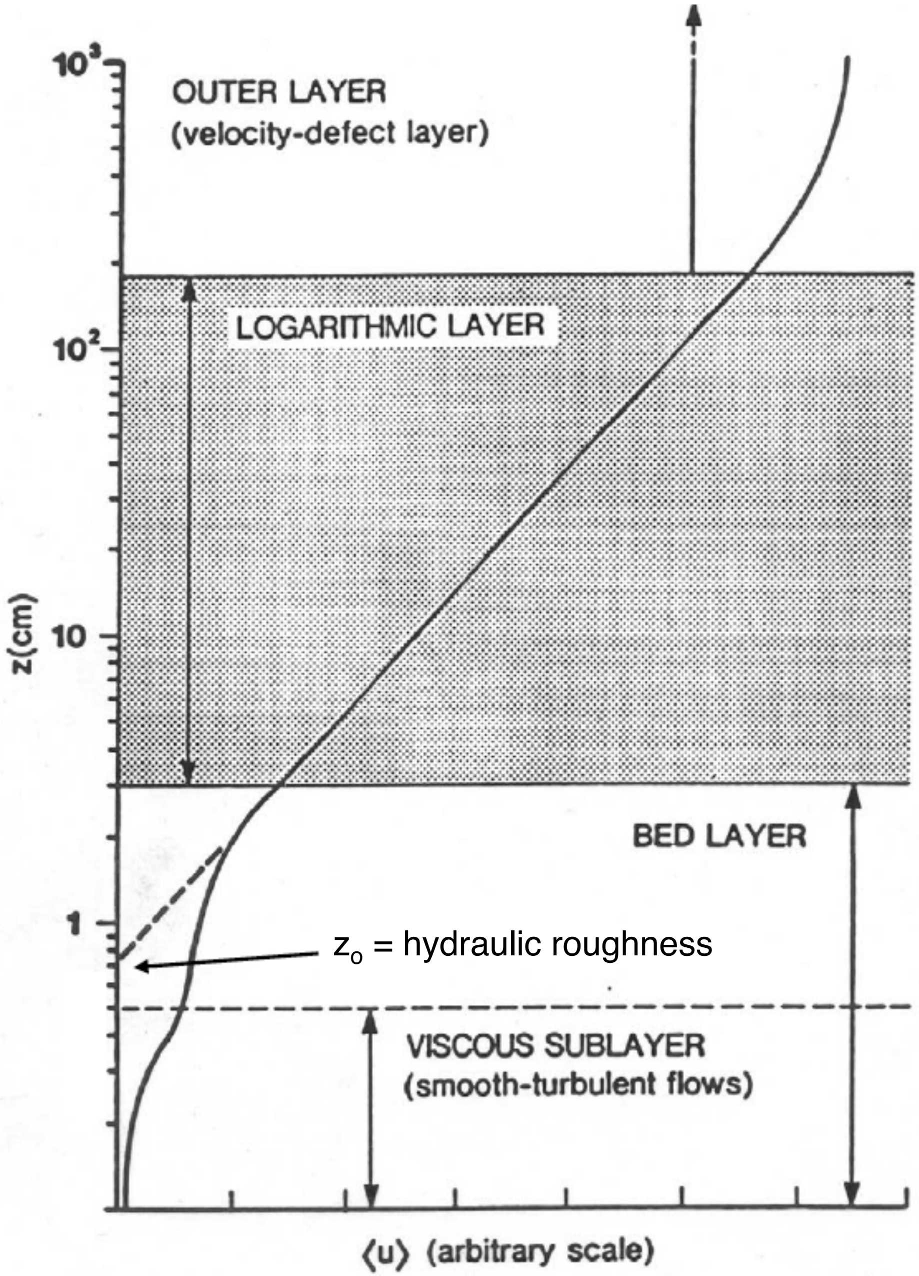
→ $$u = \frac{u_*}{\kappa}\log\left(\frac{z}{z_0}\right)$$

~ 1 cm

Viscous Layer – z/h << 1, but $\nu/(zu_*)$ not small

$$\frac{z}{u_*}\frac{du}{dz} = f\left(\frac{zu_*}{\nu}\right)$$

Bottom boundary layer often plotted on log(z) axis:



$$\frac{z}{u_*}\frac{du}{dz} = f(z/h)$$

z/h << 1

$$\frac{z}{u_*}\frac{du}{dz} = \frac{1}{\kappa}$$

"Overlap" layer

$$u = \frac{u_*}{\kappa}\log\left(\frac{z}{z_0}\right)$$

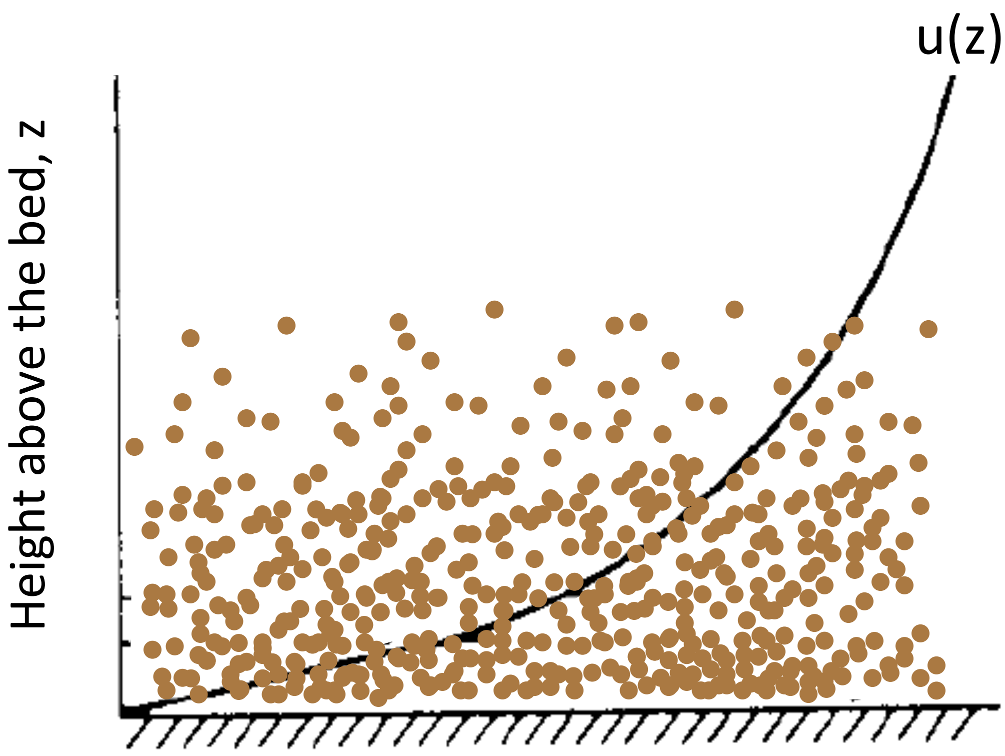
$\nu/(zu_*)$ << 1

$$\frac{z}{u_*}\frac{du}{dz} = f(zu_*/\nu)$$

(Wright, 1995)

OUTLINE: 1) Ri # importance; **2) Overlap layer**; 3) Under-saturation; 4) Critical Saturation; 5) Over-saturation

Dimensionless analysis of overlap layer with (sediment-induced) stratification:



Additional variable

b = Turbulent buoyancy flux

$$b = \frac{gs < c'w' >}{\rho_s}$$

s = $(\rho_s - \rho)/\rho \approx 1.6$

c = sediment mass conc.

w = vertical fluid vel.

$$\frac{\kappa z}{u_*}\frac{du}{dz} = 1$$

$$\rightarrow \quad \frac{\kappa z}{u_*}\frac{du}{dz} = f\left(\frac{b\kappa z}{u_*^3}\right)$$

Dimensionless ratio

$$\frac{b\kappa z}{u_*^3} \equiv \varsigma$$ = "stability parameter"

Deriving impact of z on structure of overlap (a.k.a. "log" or "wall") layer

$$\frac{\kappa z}{u_*}\frac{du}{dz} = f\left(\frac{b\kappa z}{u_*^3}\right) \longrightarrow \frac{\kappa z}{u_*}\frac{du}{dz} = f(\varsigma)$$

Rewrite f(ζ) as Taylor expansion around ζ = 0:

$$\frac{\kappa z}{u_*}\frac{du}{dz} = f(\varsigma) = \underbrace{f\Big|_{\varsigma=0}}_{=1} + \varsigma \underbrace{\frac{df}{d\varsigma}\Big|_{\varsigma=0}}_{=\alpha} + \underbrace{\frac{\varsigma^2}{2}\frac{d^2 f}{d\varsigma^2}\Big|_{\varsigma=0}}_{\approx 0} + \underbrace{\ldots}_{\approx 0}$$

$$\longrightarrow \frac{\kappa z}{u_*}\frac{du}{dz} = 1 + \alpha\varsigma \longrightarrow u = \frac{u_*}{\kappa}\left[\log\left(\frac{z}{z_0}\right) + \alpha\int_{z_0}^{z}\left(\frac{\varsigma}{z}\right)dz\right]$$

From atmospheric studies, α ≈ 4 - 5

If there is stratification (ζ > 0) then u(z) increases <u>faster</u> with ζ than homogeneous case.

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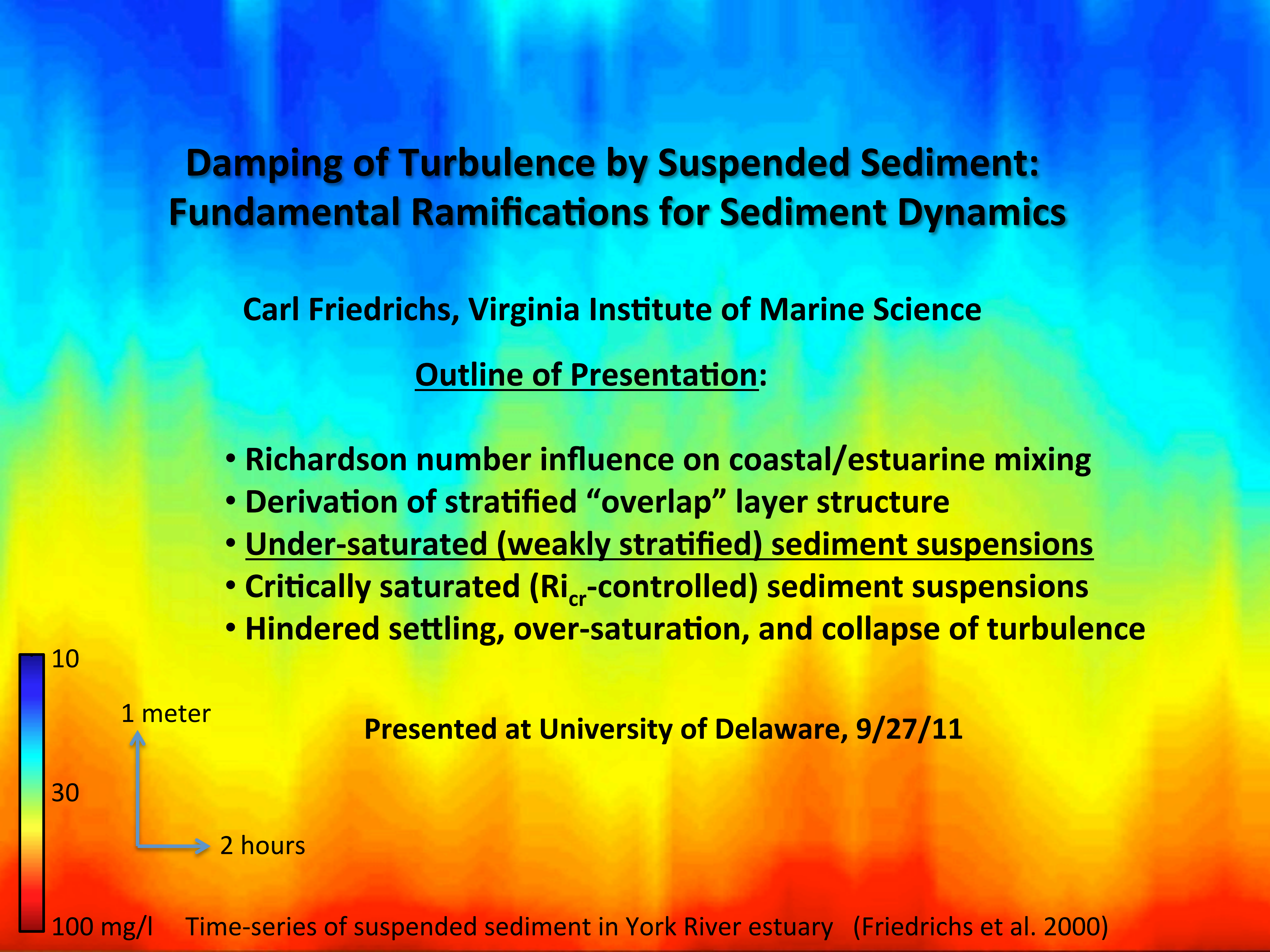


Time-series of suspended sediment in York River estuary (Friedrichs et al. 2000)

$$u = \frac{u_*}{\kappa}\left[\log\left(\frac{z}{z_0}\right) + \alpha\int_{z_0}^{z}\left(\frac{\varsigma}{z}\right)dz\right] \quad \text{Eq. (1)}$$

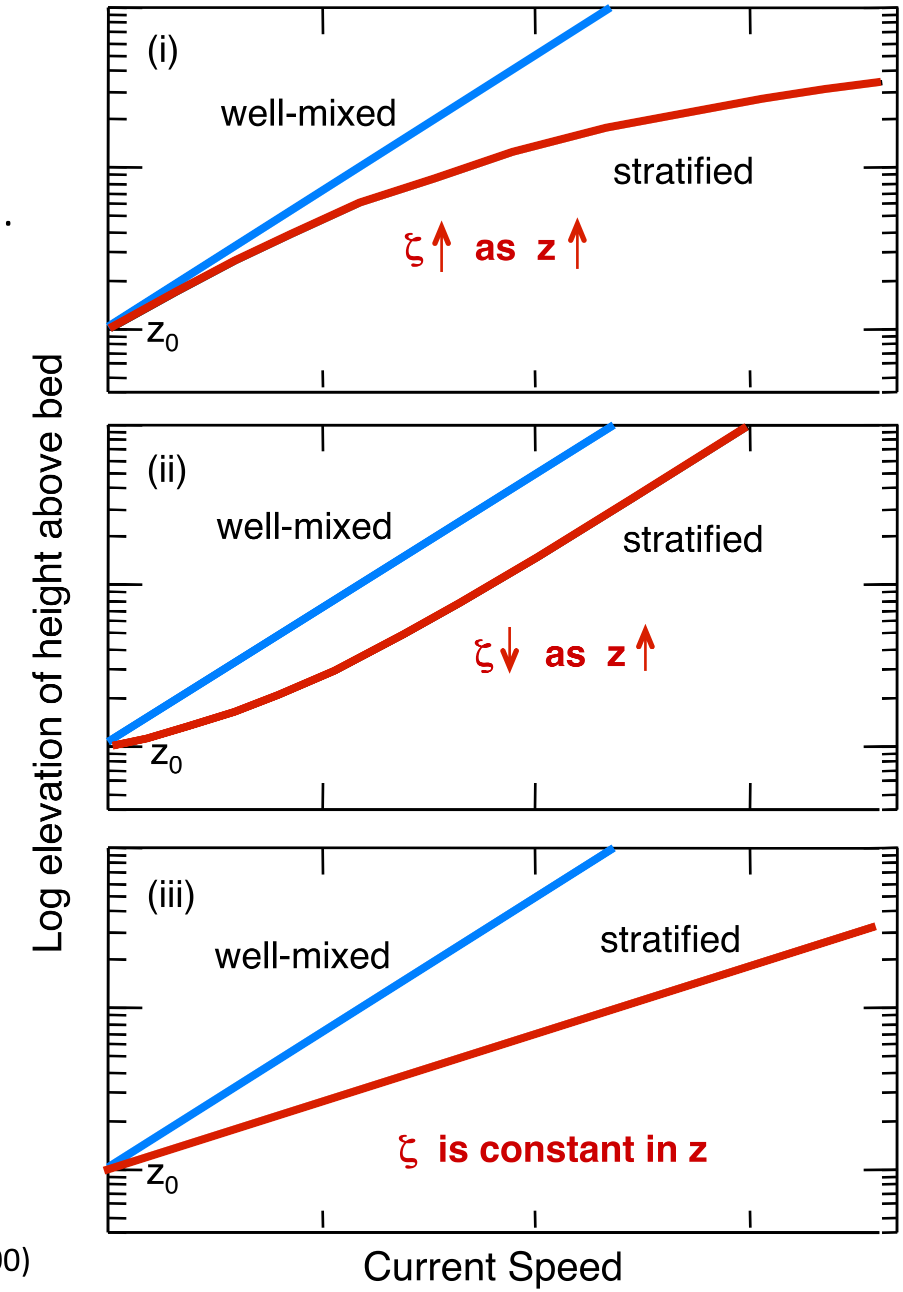
-- Case (i): No stratification near the bed ($\zeta = 0$ at $z = z_0$). Stratification and ζ increase with increased z.
-- Eq. (1) gives u increasing faster and faster with z relative to classic well-mixed log-layer.
(e.g., halocline being mixed away from below)

-- Case (ii): Stratified near the bed ($\zeta > 0$ at $z = z_0$). Stratification and ζ decreases with increased z.
-- Eq. (1) gives u initially increasing faster than u, but then matching du/dz from neutral log-layer.
(e.g., fluid mud being entrained by wind-driven flow)

-- Case (iii): uniform ζ with z. Eq (1) integrates to

$$u = \frac{u_*}{\kappa}(1+\alpha\zeta)\log\left(\frac{z}{z_0}\right)$$

-- u remains logarithmic, but shear is increased buy a factor of $(1+\alpha\zeta)$



(Friedrichs et al, 2000)

OUTLINE: 1) Ri # importance; 2) Overlap layer; **3) Under-saturation**; 4) Critical Saturation; 5) Over-saturation

Effect of stratification (via ζ) on eddy viscosity (A_z)

$$\frac{\kappa z}{u_*}\frac{du}{dz} = 1 + \alpha\varsigma$$ Overlap layer scaling modified by buoyancy flux

$$u_*^2 = A_z \frac{du}{dz}$$ Definition of eddy viscosity

Eliminate du/dz and get $$A_z = \frac{\kappa u_* z}{(1+\alpha\varsigma)}$$

-- As stratification increases (larger ζ), A_z decreases

-- If ζ = const. in z, A_z increases like $u_* z$, and the result is still a log-profile.

Connect stability parameter, ζ, to shape of concentration profile, c(z):

Definition of ζ: $$\varsigma = \frac{b\kappa z}{u_*^3} = \frac{gs \langle c'w' \rangle \kappa z}{\rho_s u_*^3}$$

Rouse balance (Reynolds flux = settling): $$\frac{gs \langle c'w' \rangle \kappa z}{u_*^3} = c\, w_s$$

Combine to eliminate <c’w’> : $$\varsigma = \left(\frac{g s w_s \kappa}{\rho_s u_*^3}\right) c\, z$$

ζ = const. in z if $c \sim z^{-1}$

(Assuming w_s is const. in z)

$$\varsigma = \left(\frac{g s w_s \kappa}{\rho_s u_*^3} \right) c\, z \qquad \zeta = \text{const. in } z \quad \text{if} \quad c \sim z^{-1}$$

Fit a general power-law to c(z) of the form $c \sim z^{-A}$

Then $\varsigma \sim z^{(1-A)}$

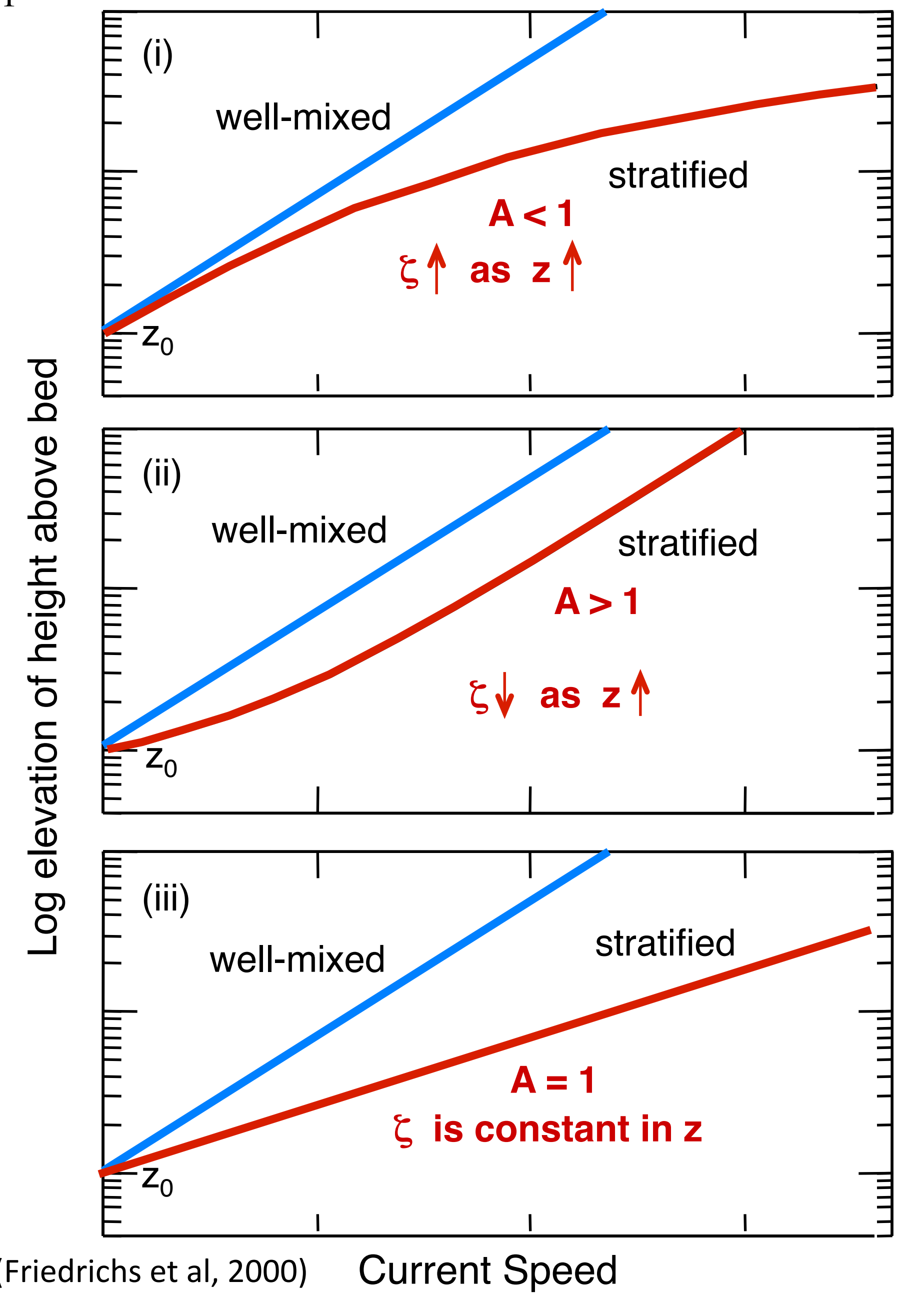
If A < 1, c decreases more slowly than z^{-1}
ζ increases with z,
stability increases upward,
u is more concave-down than log(z)

If A > 1, c increases more quickly than z^{-1}
ζ decreases with z, stability
becomes less pronounced upward,
u is more concave-up than log(z)

If A = 1, $c \sim z^{-1}$
ζ is constant with elevation
stability is uniform in z,
u follows log(z) profile

If suspended sediment concentration, $C \sim z^{-A}$

Then A <,>,= 1 determines shape of u profile



(Friedrichs et al, 2000)

If suspended sediment concentration, $C \sim z^{-A}$

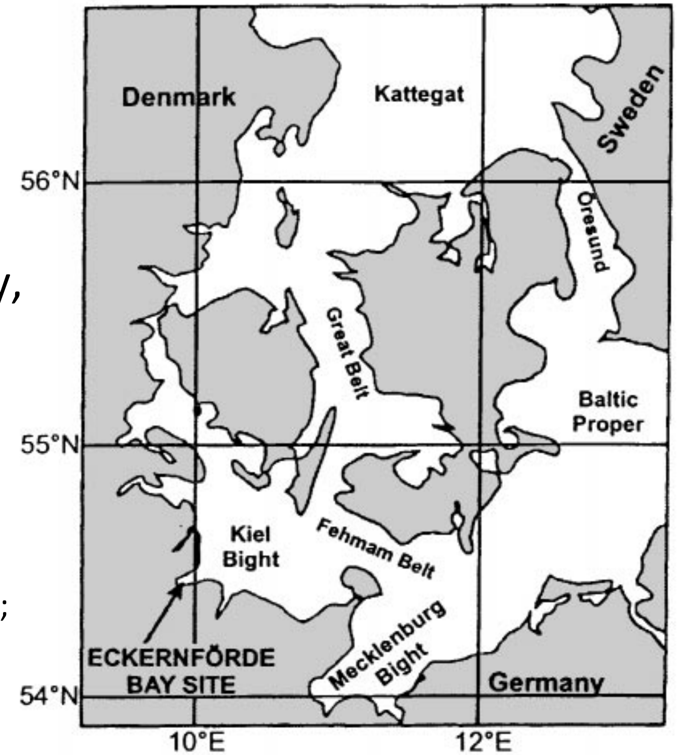
A < 1 predicts u more concave-down than log(z)

A > 1 predicts u more concave-up than log(z)

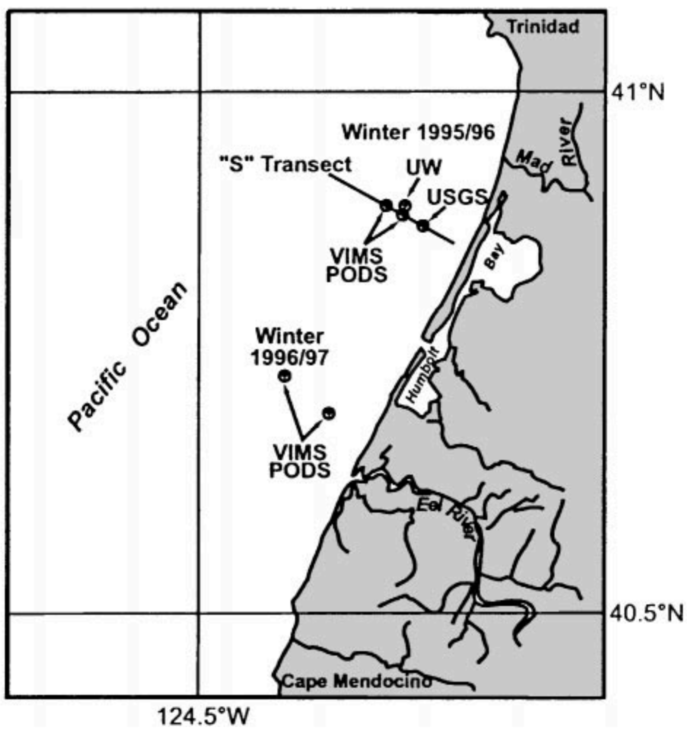
A = 1 predicts u will follow log(z)

Testing this relationship using observations from bottom boundary layers:

Eckernförde Bay, Baltic Coast, Germany

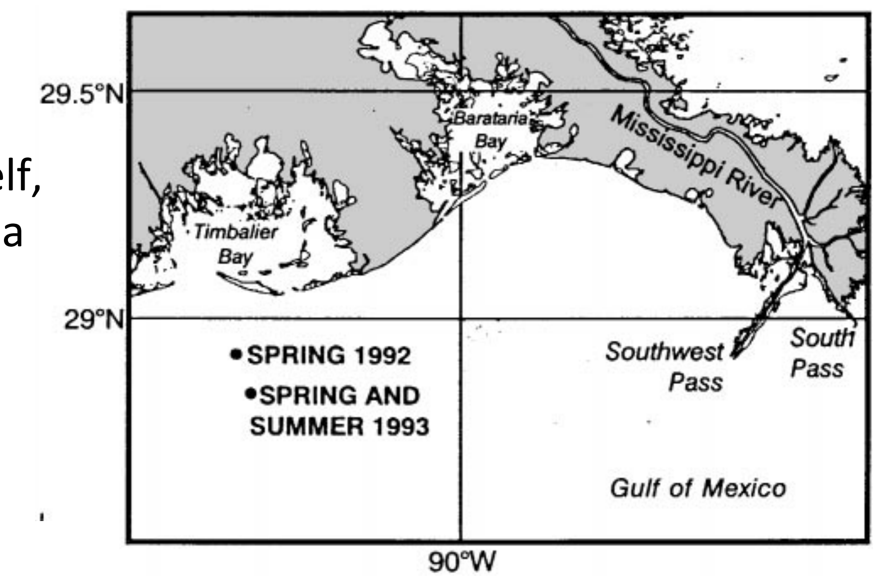


STRATAFORM mid-shelf site, Northern California, USA



(Friedrichs & Wright, 1997; Friedrichs et al, 2000)

Inner shelf, Louisiana USA



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If suspended sediment concentration, $C \sim z^{-A}$

A < 1 predicts u more concave-down than log(z)

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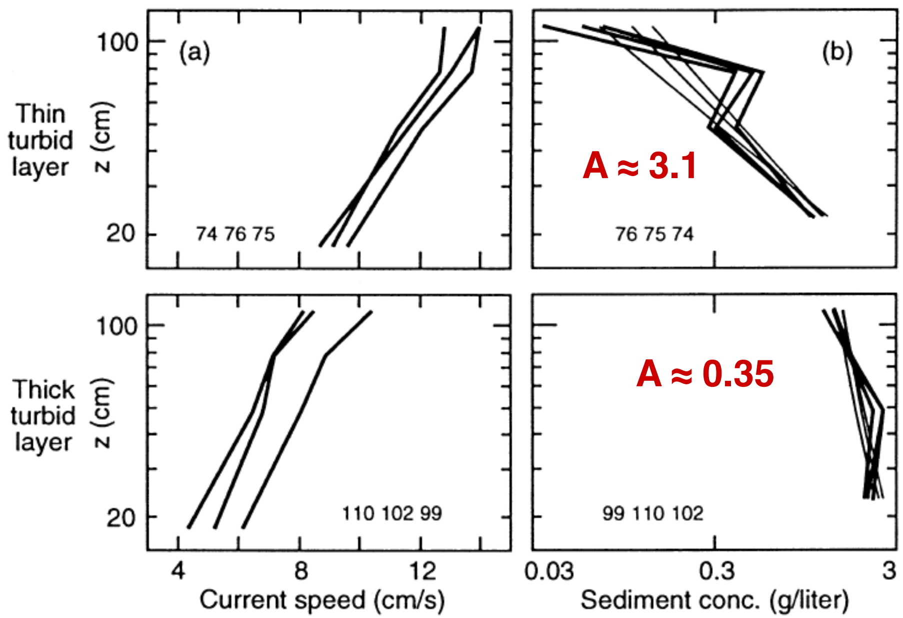
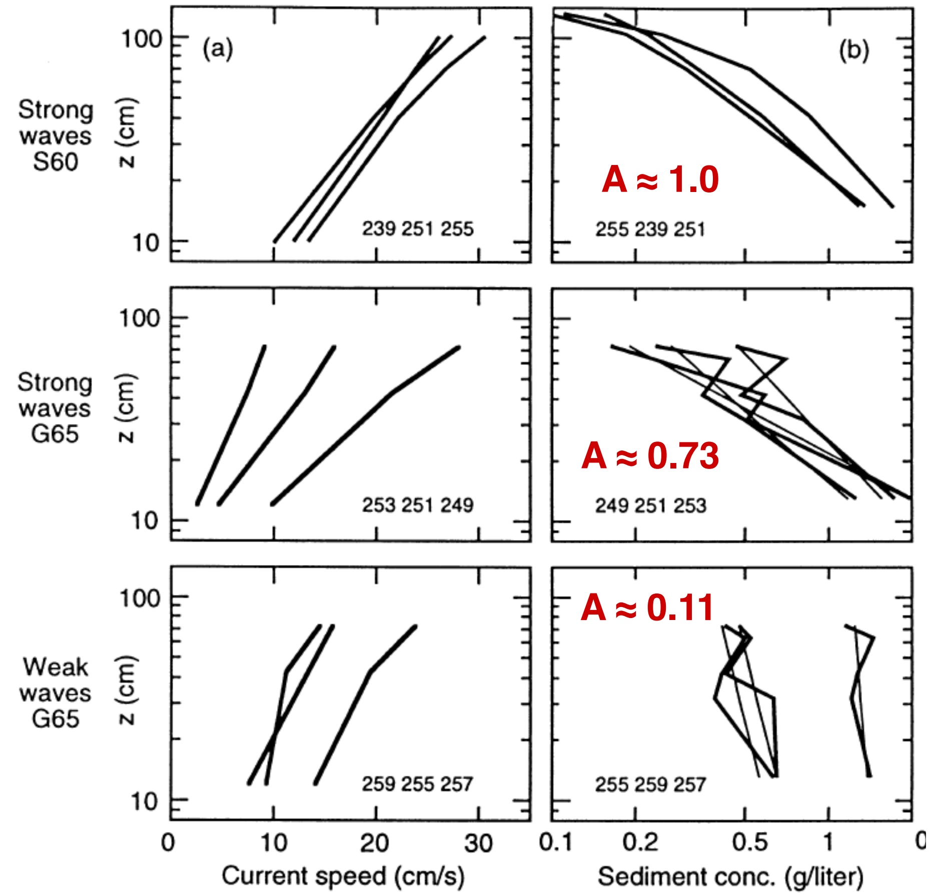
A = 1 predicts u will follow log(z)

Inner shelf, Louisiana, USA, 1993



(Friedrichs et al, 2000)

STATAFORM mid-shelf site, Northern California, USA, 1995, 1996

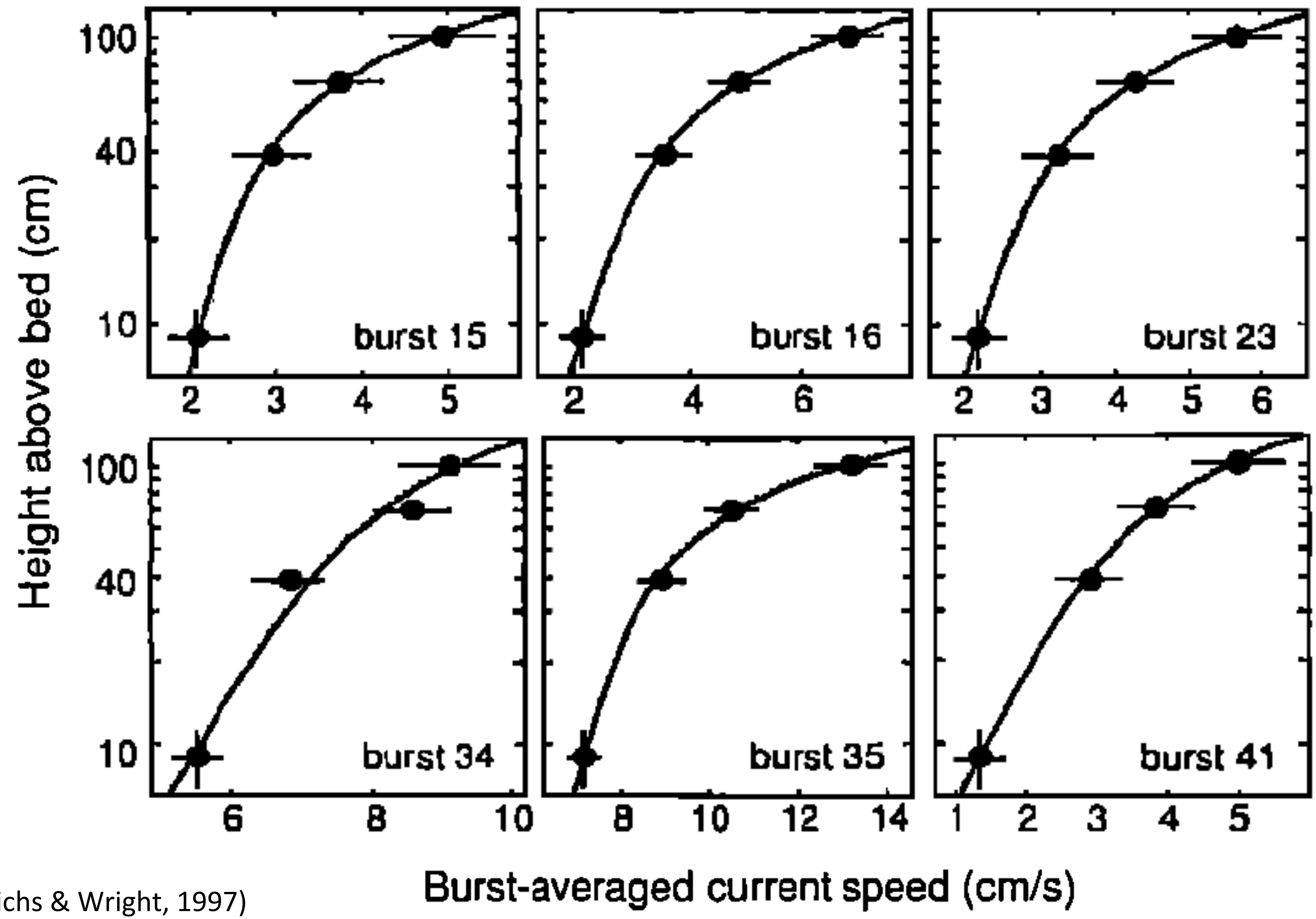


-- Smallest values of A < 1 are associated with concave-downward velocities on log-plot.

-- Largest value of A > 1 is associated with concave-upward velocities on log-plot.

-- Intermediate values of A ≈ 1 are associated with straightest velocities on log-plot.

Eckernförde Bay, Baltic Coast, Germany, spring 1993

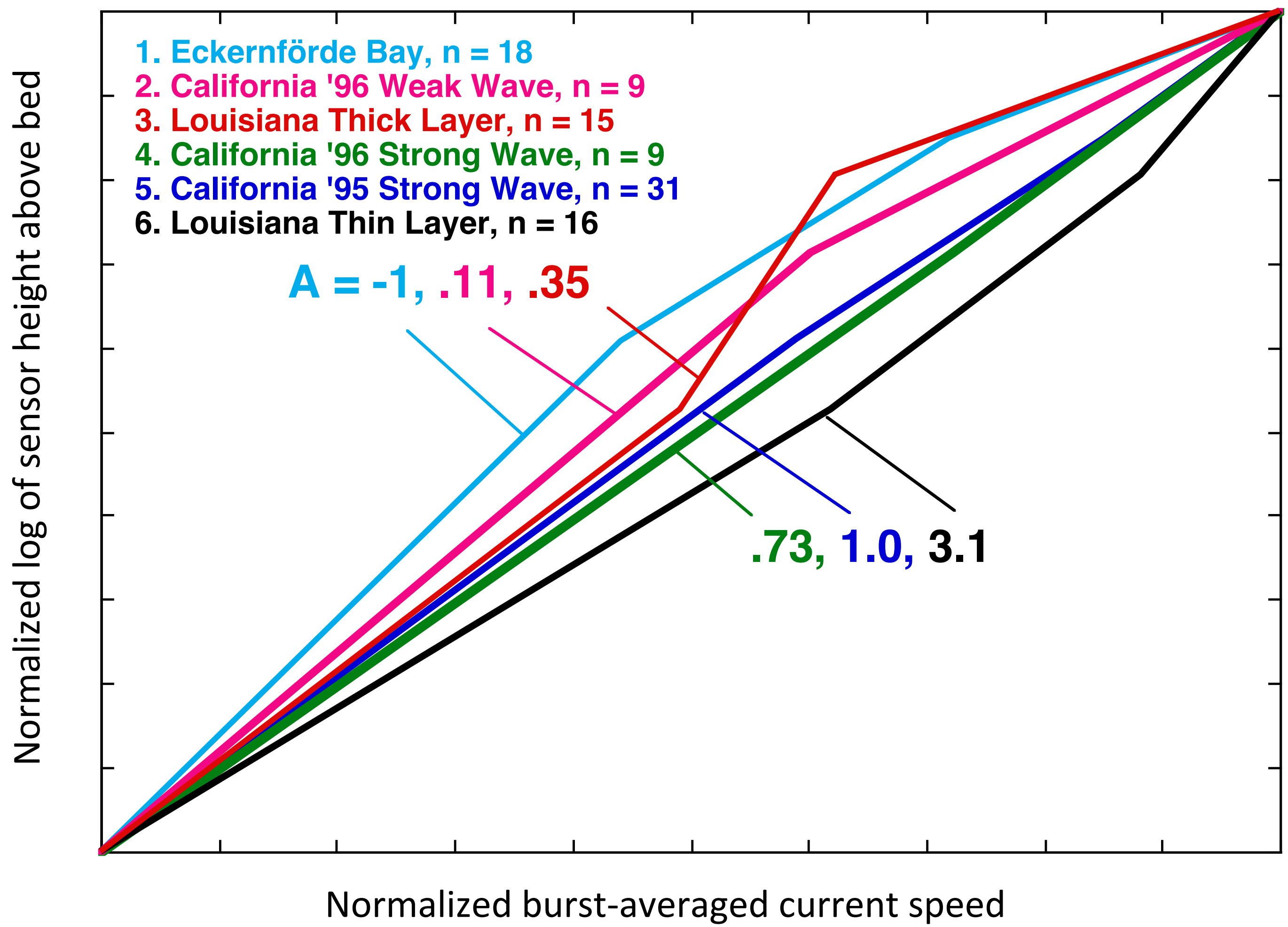


(Friedrichs & Wright, 1997)

-- Salinity stratification that increases upwards cannot be directly represented by c ~ z^{-A}. Friedrichs et al. (2000) argued that this case is dynamically analogous to A ≈ -1.

OUTLINE: 1) Ri # importance; 2) Overlap layer; **3) Under-saturation**; 4) Critical Saturation; 5) Over-saturation

Observations showing effect of concentration exponent A on shape of velocity profile



Observations also show:

A < 1, concave-down velocity

A > 1, concave-up velocity

A ~ 1, straight velocity profile

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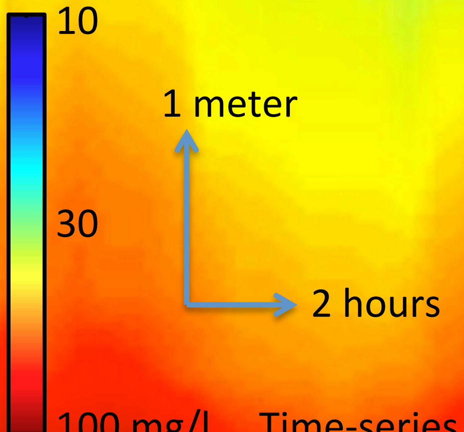


Time-series of suspended sediment in York River estuary (Friedrichs et al. 2000)

Relate stability parameter, ζ, to Richardson number:

Definition of gradient Richardson number associated with suspended sediment:

$$Ri = -\frac{gs(dc/dz)}{\rho_s (du/dz)^2}$$

Original definition and application of ζ:

$$\varsigma = \frac{gs <c'w'> \kappa z}{\rho_s u_*^3} \qquad \frac{\kappa z}{u_*}\frac{du}{dz} = 1 + \alpha\varsigma$$

Relation found for eddy viscosity:

$$A_z = \frac{\kappa u_* z}{(1+\alpha\varsigma)}$$

Definition of eddy diffusivity:

$$-<c'w'> = K_z \frac{dc}{dz}$$

Assume momentum and mass are mixed similarly: $A_z = K_z$

Combine all these and you get:

$$Ri = \frac{\varsigma}{1+\alpha\varsigma}$$

So a constant ζ with height also leads to a constant Ri with height.

Also, if ζ increases (or decreases) with height Ri correspondingly increases (or decreases).

$$\varsigma = \left(\frac{g s w_s \kappa}{\rho_s u_*^3} \right) c\, z \qquad Ri = \frac{\varsigma}{1+\alpha\varsigma}$$

ζ and Ri const. in z if $c \sim z^{-1}$

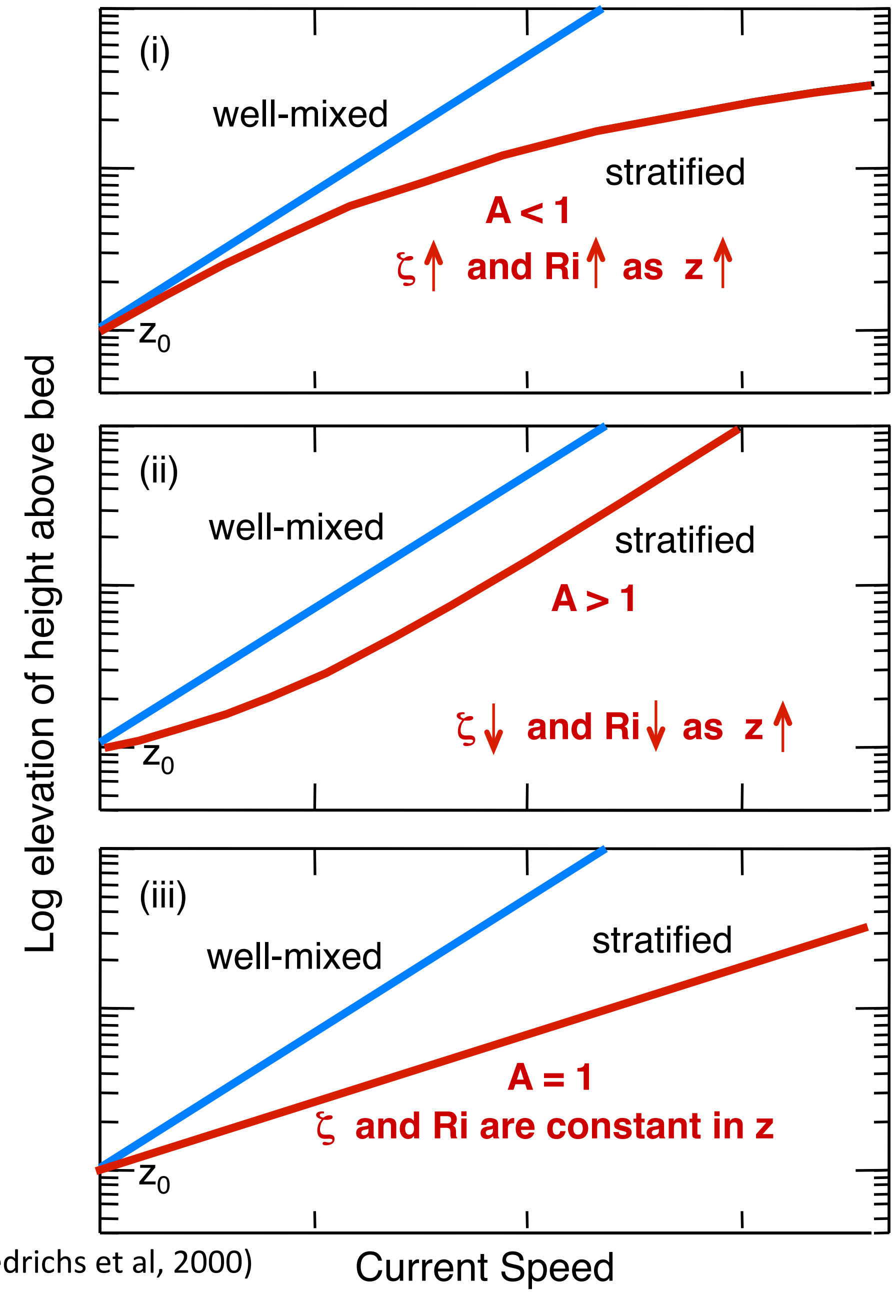
Define $c \sim z^{-A}$ then $\varsigma \sim z^{(1-A)}$

If A < 1, c decreases more slowly than z^{-1}
ζ and Ri increase with z,
stability increases upward,
u is more concave-down than log(z)

If A > 1, c decreases more quickly than z^{-1}
ζ and Ri decrease with z, stability
becomes less pronounced upward,
u is more concave-up than log(z)

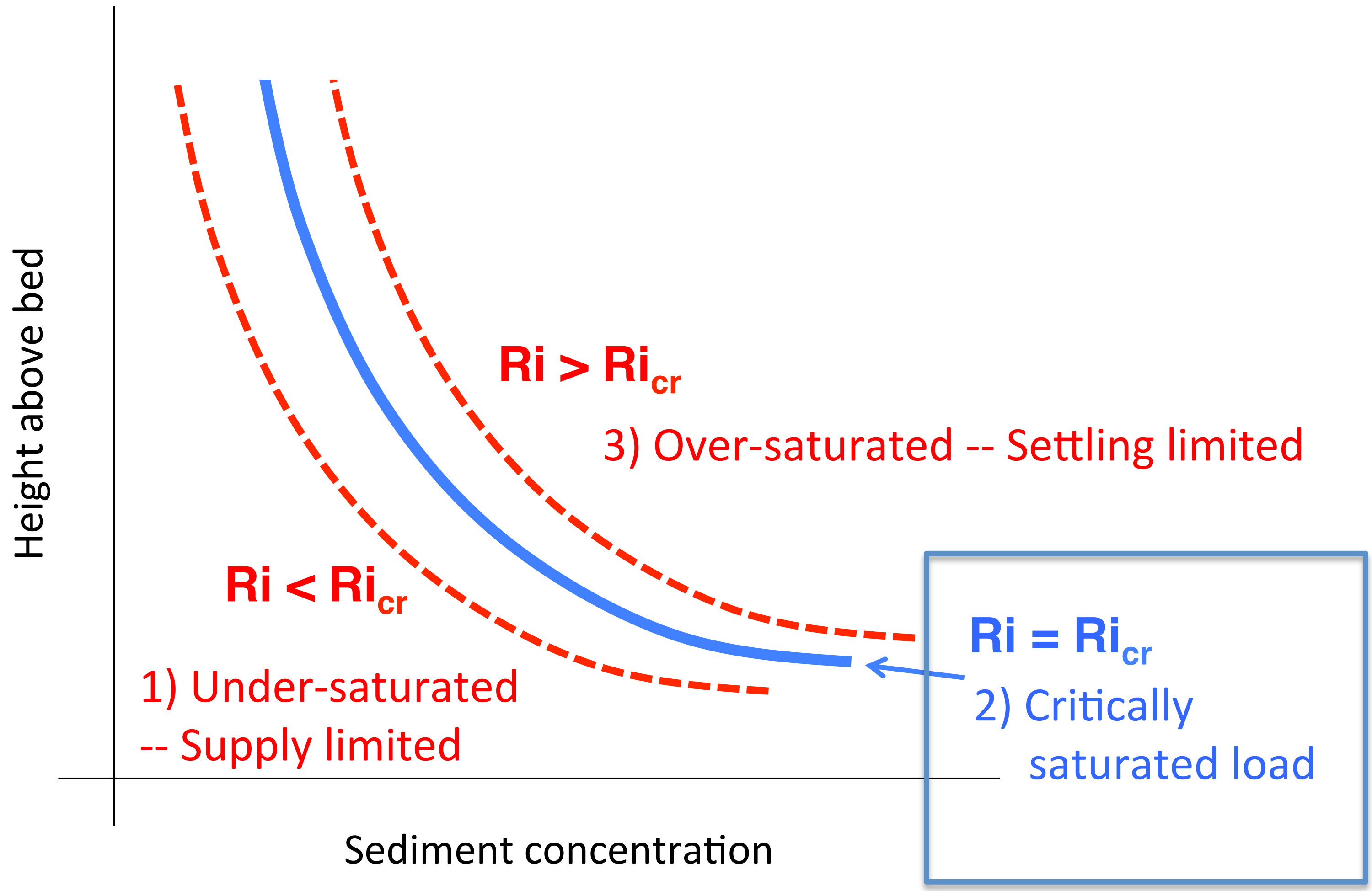
If A = 1, $c \sim z^{-1}$
ζ and Ri are constant with elevation
stability is uniform in z,
u follows log(z) profile

If suspended sediment concentration, $C \sim z^{-A}$
then A <,>,= 1 determines shape of u profile
and also the vertical trend in ζ and Ri



(Friedrichs et al, 2000)

Now focus on the case where $Ri = Ri_{cr}$ (so Ri is constant in z over "log" layer)



Connection between structure of sediment settling velocity to structure of "log-layer" when $Ri = Ri_{cr}$ in z (and therefore ζ is constant in z too).

Rouse Balance: $$w_s C = K_z \frac{dc}{dz}$$

Earlier relation for eddy viscosity: $$K_z = \frac{\kappa u_* z}{(1+\alpha\varsigma)}$$

Eliminate K_z and integrate in z to get

$$\frac{C}{C_{ref}} = \left(\frac{z}{z_{ref}}\right)^{-\left[\frac{w_s(1+\alpha\varsigma)}{\kappa u_*}\right]}$$

But we already know $c \sim z^{-1}$ when Ri = const.

So $$\frac{w_s(1+\alpha\varsigma)}{\kappa u_*} = 1$$ and $$1+\alpha\varsigma = \frac{\kappa u_*}{w_s}$$ when $Ri = Ri_{cr}$

$$1+\alpha\varsigma = \frac{\kappa u_*}{w_s}$$ when Ri = Ri_{cr} . This also means that when Ri = Ri_{cr} :

$$Ri = \frac{\varsigma}{1+\alpha\varsigma} \longrightarrow Ri_{cr} = \frac{w_s\,\varsigma}{\kappa u_*}$$

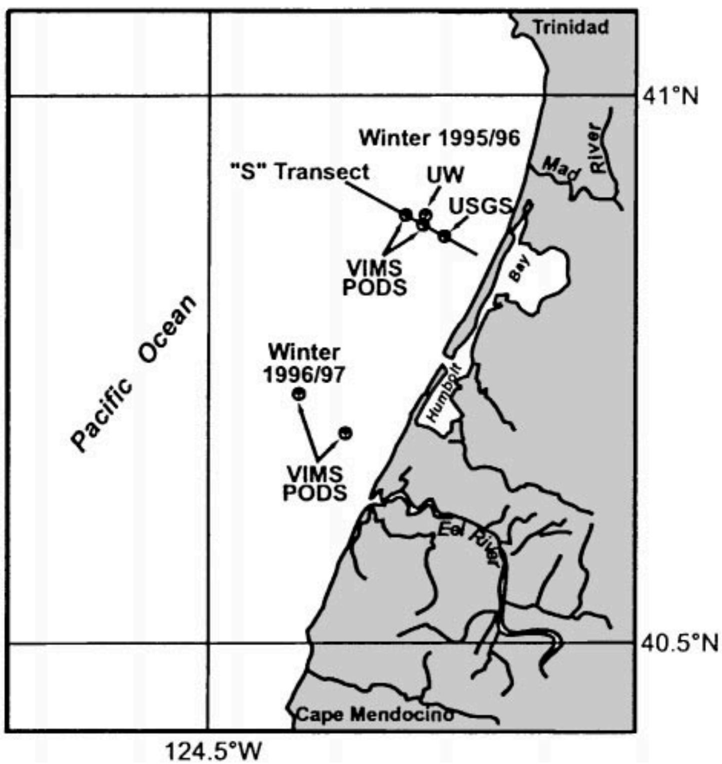
$$A_z = K_z = \frac{\kappa u_* z}{(1+\alpha\varsigma)} \longrightarrow A_z = K_z = w_s z$$

$$\frac{du}{dz} = \frac{u_*}{\kappa z}(1+\alpha\varsigma) \longrightarrow \frac{du}{dz} = \frac{u_*^2}{w_s z}$$

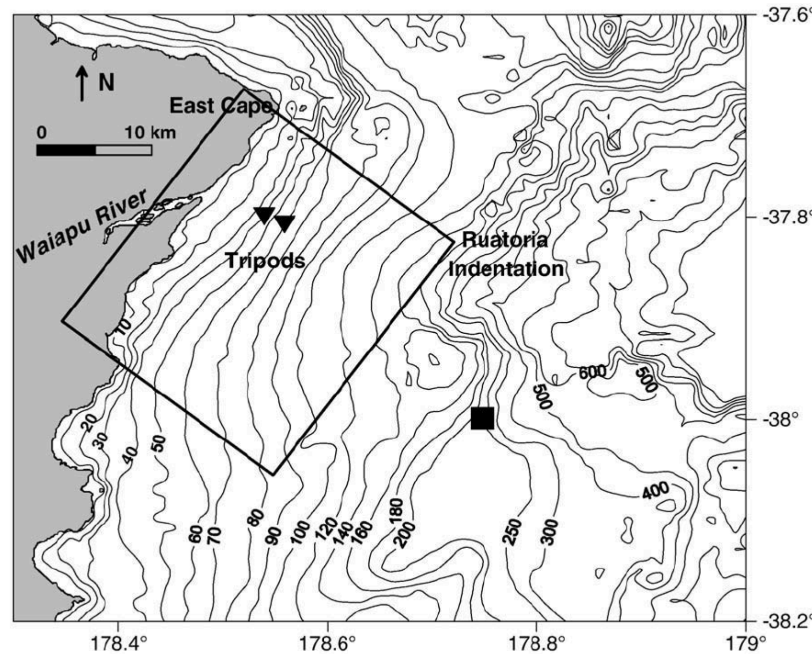
$$u = \frac{u_*}{\kappa}(1+\alpha\zeta)\log\left(\frac{z}{z_0}\right) \longrightarrow u = \frac{u_*^2}{w_s}\log\left(\frac{z}{z_0}\right)$$

$$Ri = -\frac{gs(dc/dz)}{\rho_s(du/dz)^2} \longrightarrow c = \frac{Ri_{cr}\,\rho_s}{gs}\left(\frac{u_*^2}{w_s}\right)^2 z^{-1}$$

STATAFORM mid-shelf site,
Northern California, USA



Mid-shelf site off
Waiapu River, New Zealand



(Wright, Friedrichs et al., 1999;
Maa, Friedrichs, et al., 2010)

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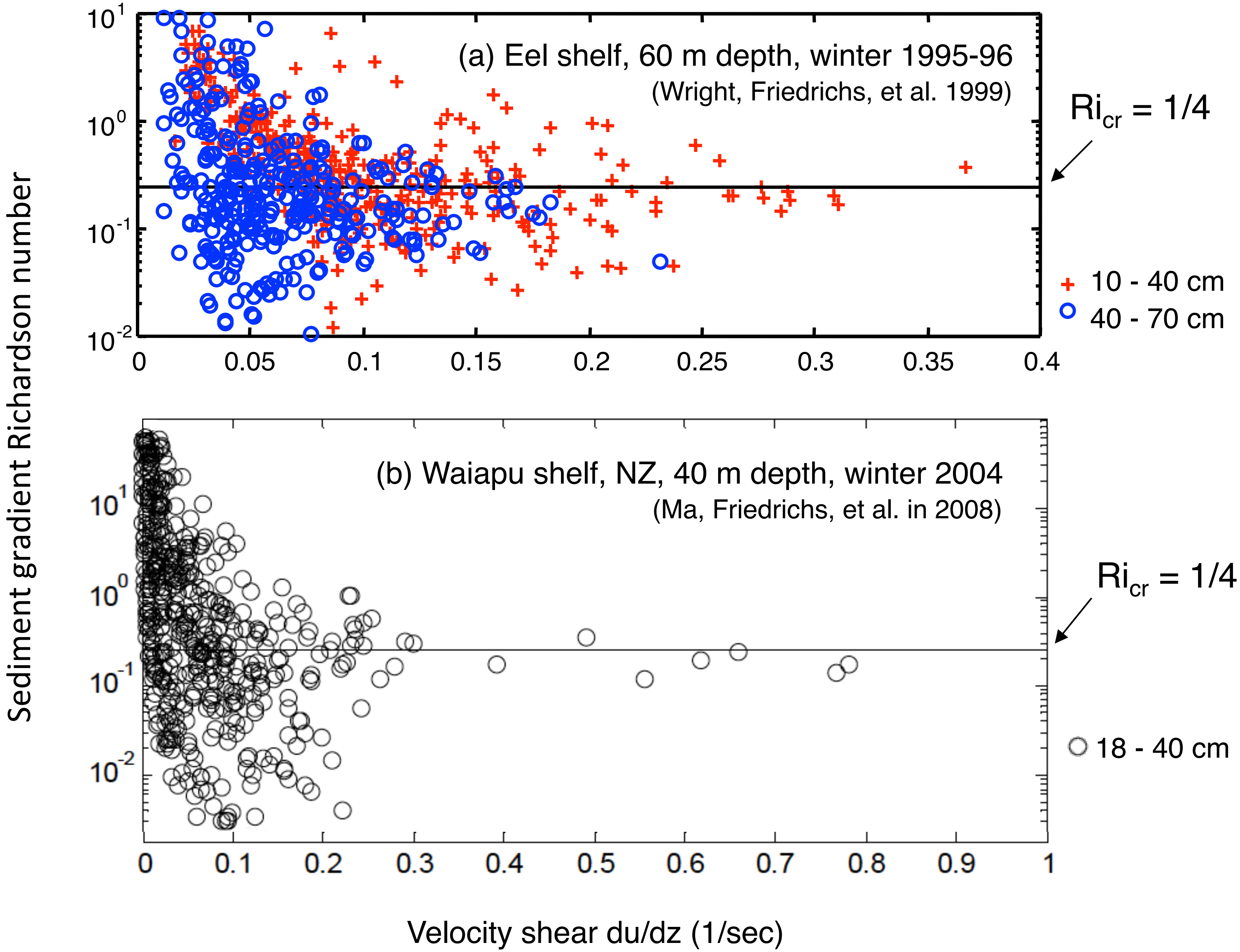
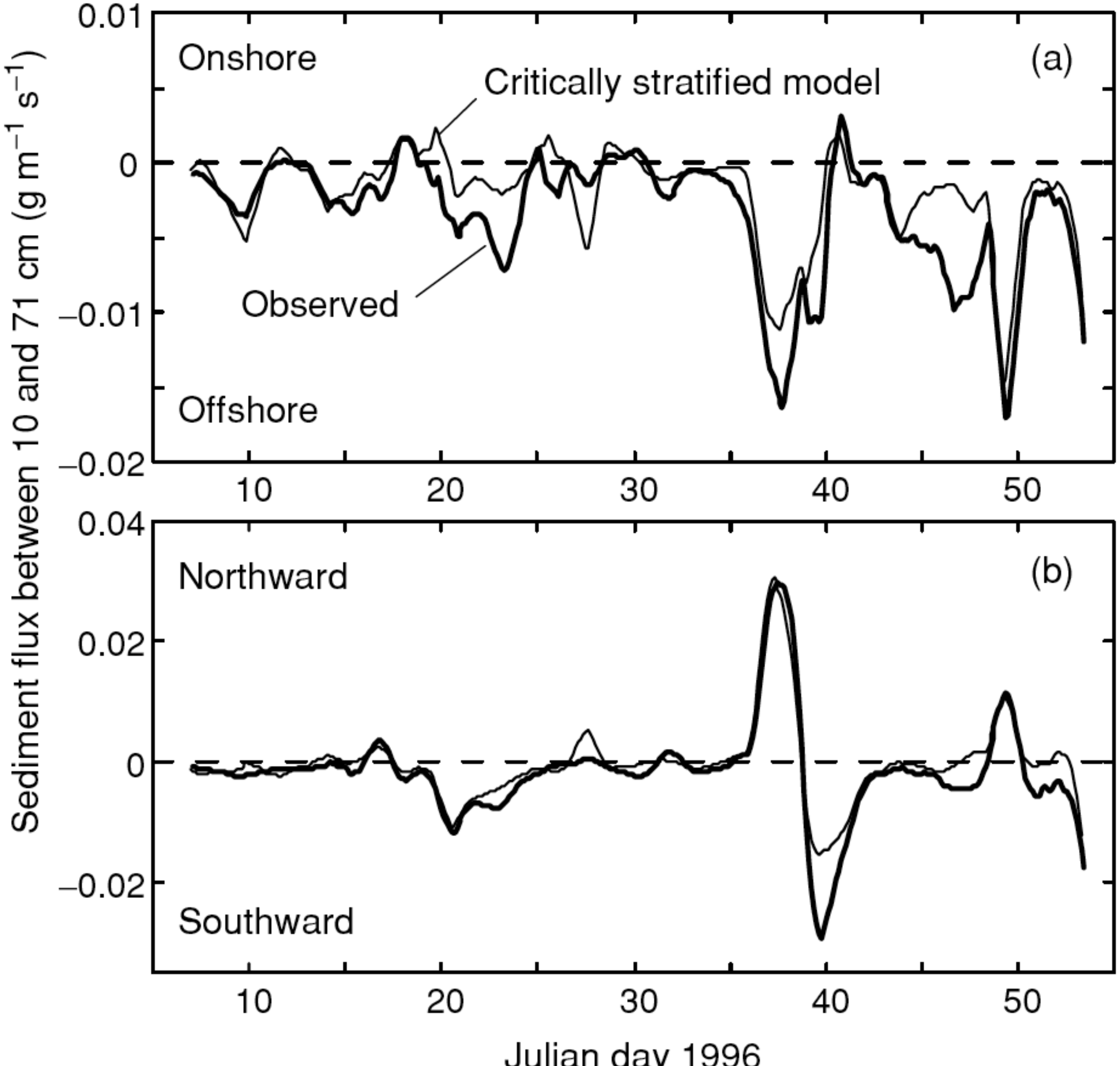


OUTLINE: 1) Ri # importance; 2) Overlap layer; 3) Under-saturation; **4) Critical Saturation**; 5) Over-saturation

Application of Ri_{cr} log-layer equations fo Eel shelf, 60 m depth, winter 1995-96



$$u = \frac{u_*^2}{w_s} \log\left(\frac{z}{z_0}\right)$$

$$c = \frac{Ri_{cr}\,\rho_s}{gs}\left(\frac{u_*^2}{w_s}\right)^2 z^{-1}$$

(Souza & Friedrichs, 2005)

OUTLINE: 1) Ri # importance; 2) Overlap layer; 3) Under-saturation; **<u>4) Critical Saturation</u>**; 5) Over-saturation

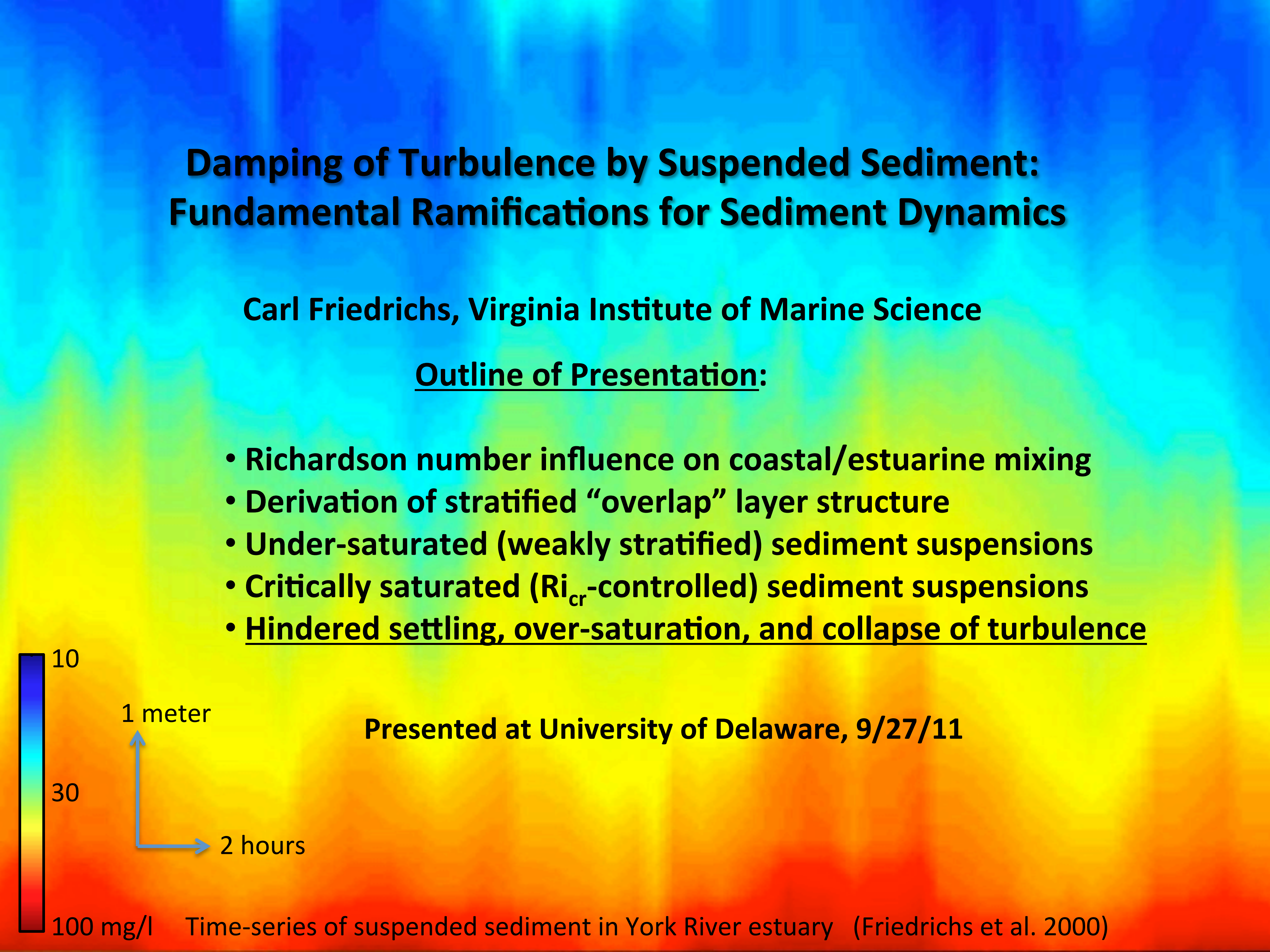
Damping of Turbulence by Suspended Sediment: Fundamental Ramifications for Sediment Dynamics

Carl Friedrichs, Virginia Institute of Marine Science

<u>Outline of Presentation</u>:

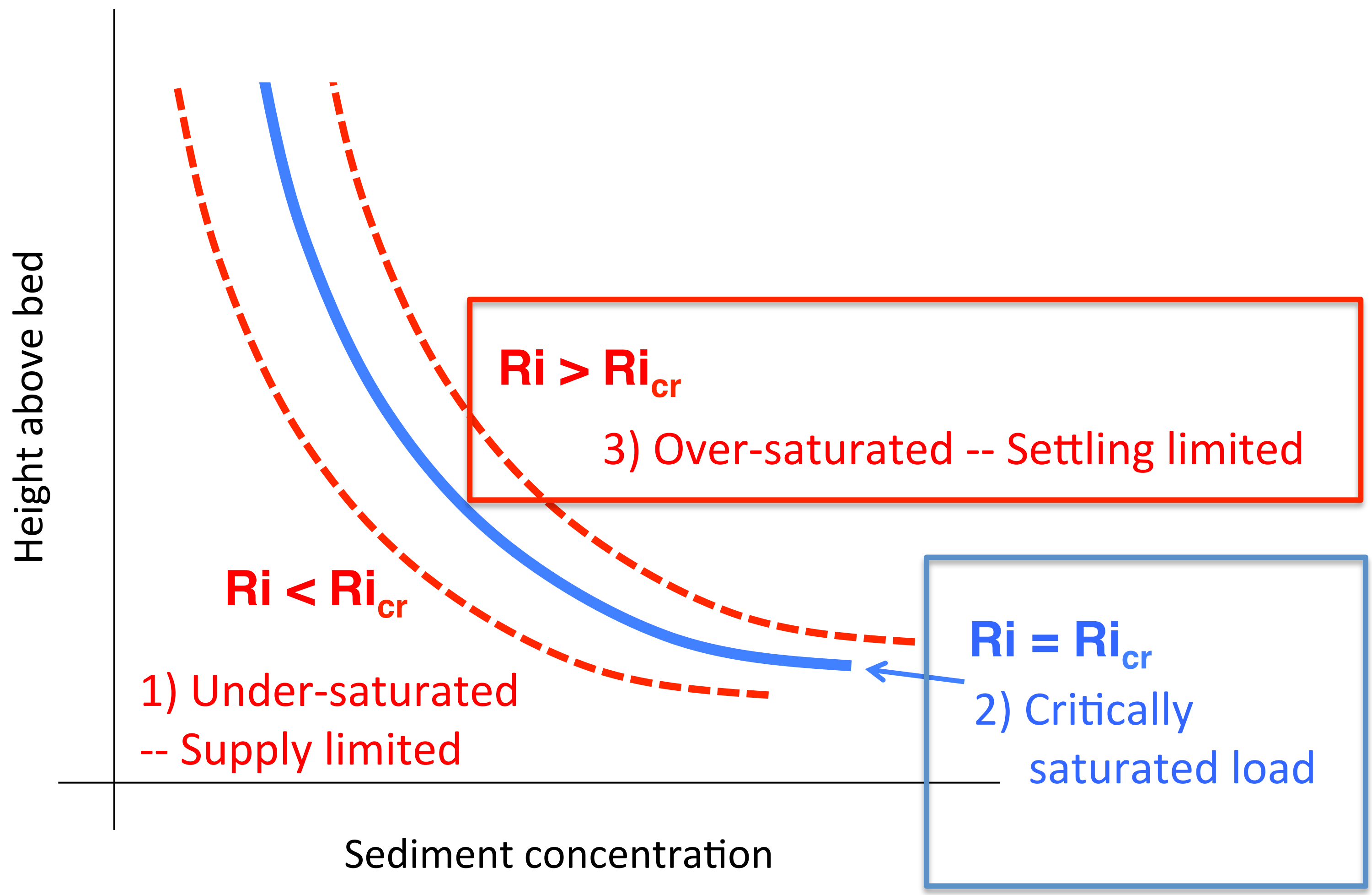
- **Richardson number influence on coastal/estuarine mixing**
- **Derivation of stratified "overlap" layer structure**
- **Under-saturated (weakly stratified) sediment suspensions**
- **Critically saturated (Ri_{cr}-controlled) sediment suspensions**
- **<u>Hindered settling, over-saturation, and collapse of turbulence</u>**

Presented at University of Delaware, 9/27/11



Time-series of suspended sediment in York River estuary (Friedrichs et al. 2000)

Now also consider over-saturated cases:



(Mehta & McAnally, 2008)

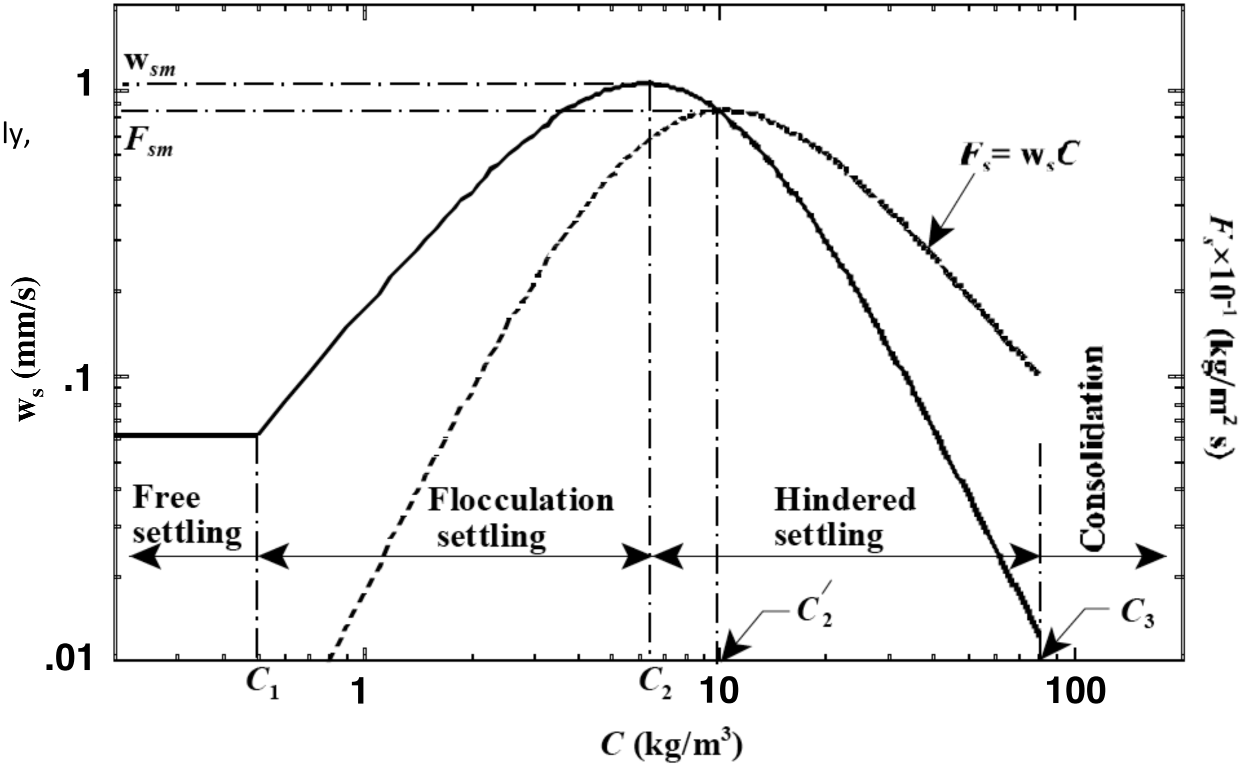


Figure 4.12 A representative plot of settling velocity and associated settling flux variation with suspension concentration.

Starting at around 5 - 8 grams/liter, the return flow of water around settling flocs creates so much drag on neighboring flocs that w_s starts to decrease with additional increases in concentration.

At ~ 10 g/l, w_s decreases so much with increased C that the rate of settling flux decreases with further increases in C. This is "hindered settling" and can cause a strong lutecline (vertical sediment gradient) to form.

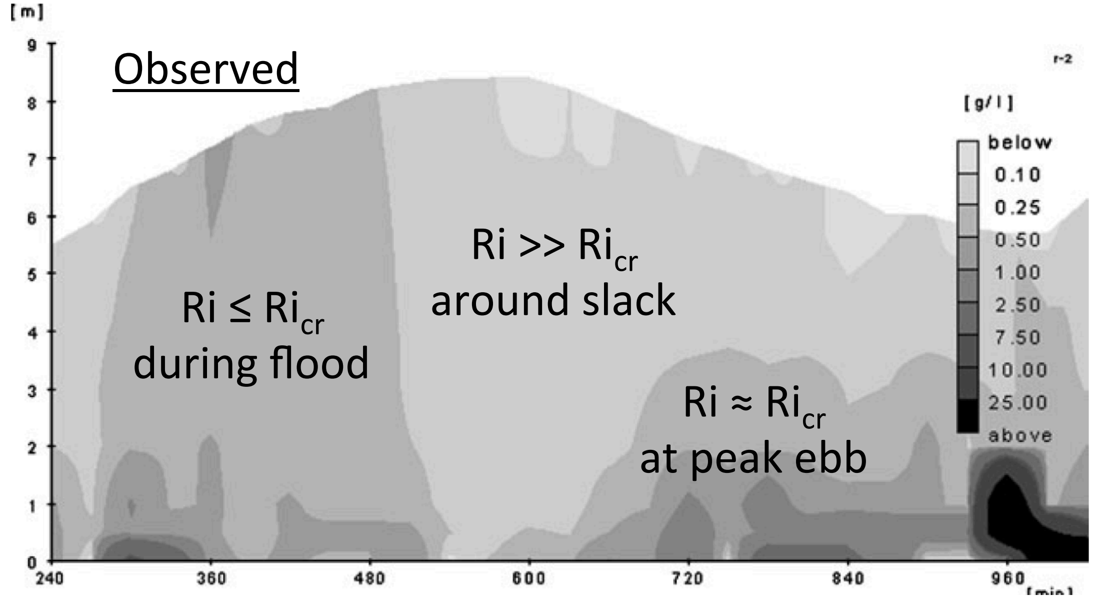
A lutecline with hindered settling can cause turbulent collapse. The sediment can't leave the water column, so dC/dz keeps increasing, creating <u>positive</u> feedback. Ri increases further above Ri_{cr}, and more sediment to settles. Then there is more hindered settling and a stronger lutecline, increases Ri further.

OUTLINE: 1) Ri # importance; 2) Overlap layer; 3) Under-saturation; 4) Critical Saturation; **5) Over-saturation**

Fine sediment transport by tidal asymmetry in the high-concentrated Ems River: indications for a regime shift in response to channel deepening

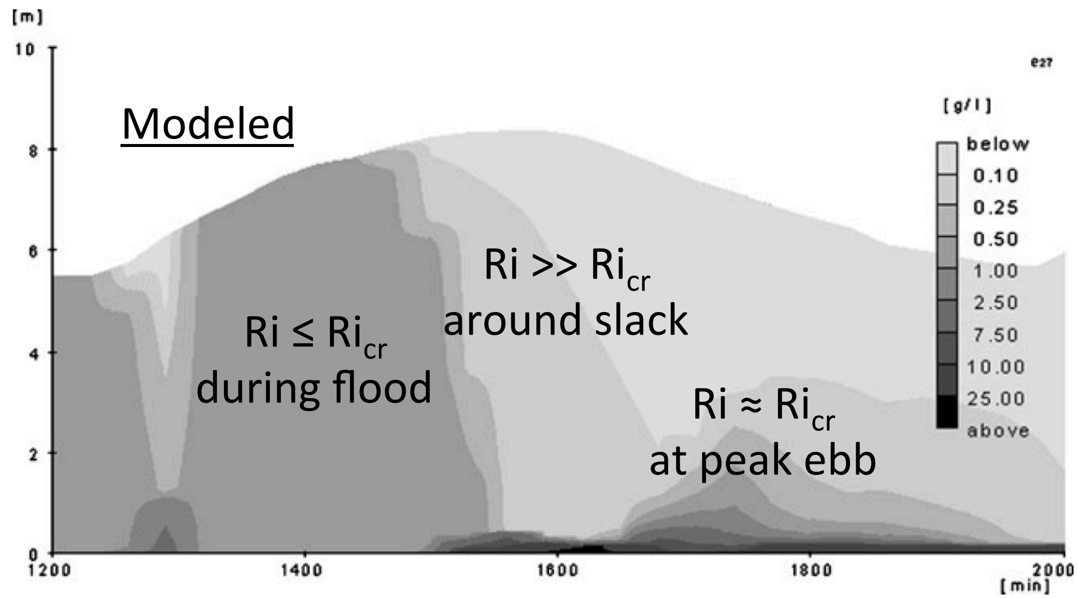
(Winterwerp, 2011)

Fig. 6 Measured isolutals at Station 2, June 19, 1990. Note rapid settling just prior to high water and pronounced stratification during ebb (after Van Leussen 1994)



$$Ri = \frac{g\ s\ \partial c/\partial z}{\rho_s\ (\partial u/\partial z)^2}$$

Fig. 7 Computed isolutals at Station 2, June 19, 1990. Note rapid settling just prior to high water and pronounced stratification during ebb



-- 1-DV k-ε model based on components of Delft 3D
-- Sediment in density formulation
-- Flocculation model
-- Hindered settling model

OUTLINE: 1) Ri # importance; 2) Overlap layer; 3) Under-saturation; 4) Critical Saturation; **5) Over-saturation**

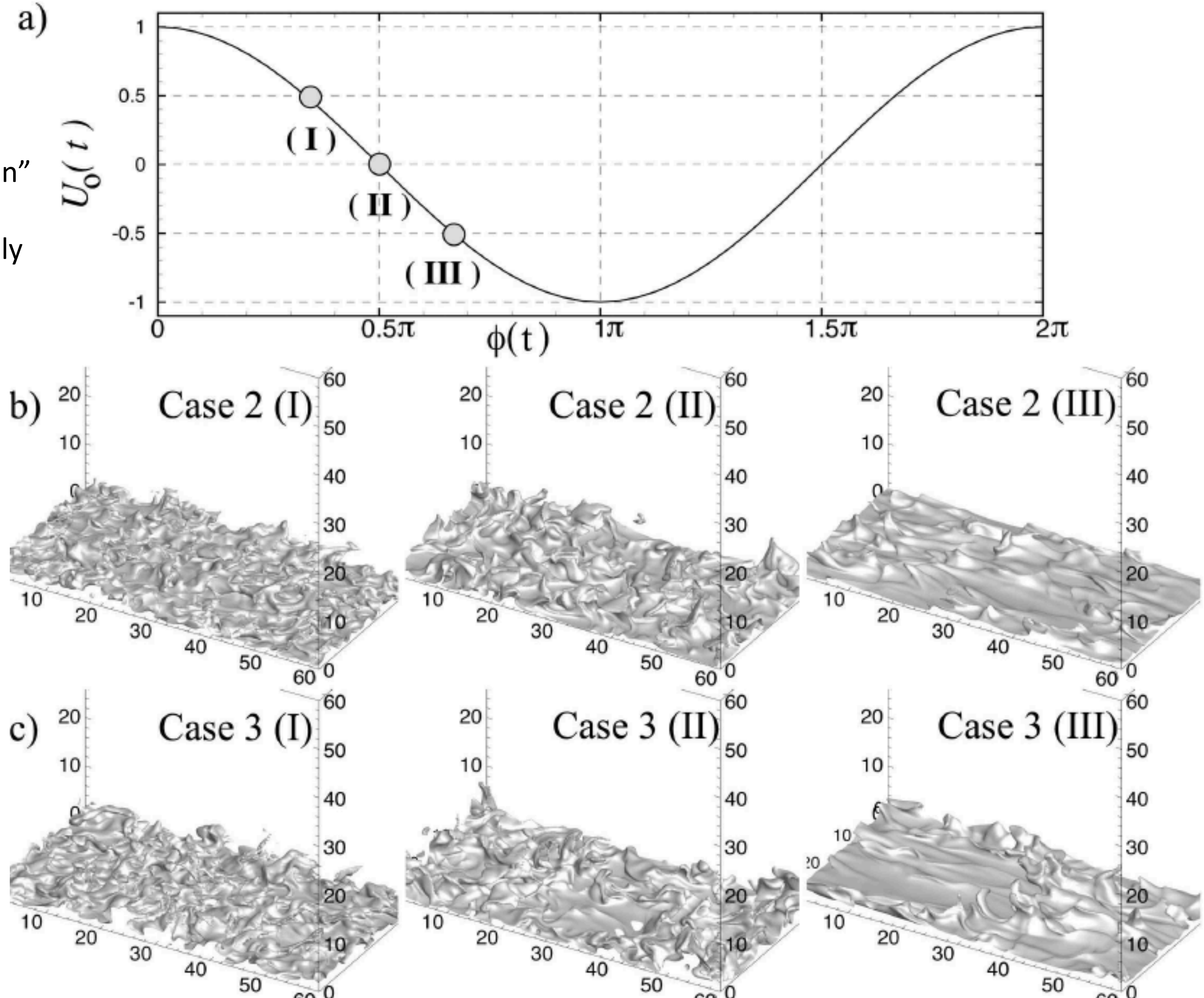
A numerical investigation of lutocline dynamics and saturation of fine sediment in the oscillatory boundary layer

(Ozedemir, Hsu & Balachandar, in press)

U ~ 60 cm/s
C ~ 10 g/liter
“large eddy simulation” model
Fixed sediment supply

w_s = 0.45 mm/s

w_s = 0.75 mm/s

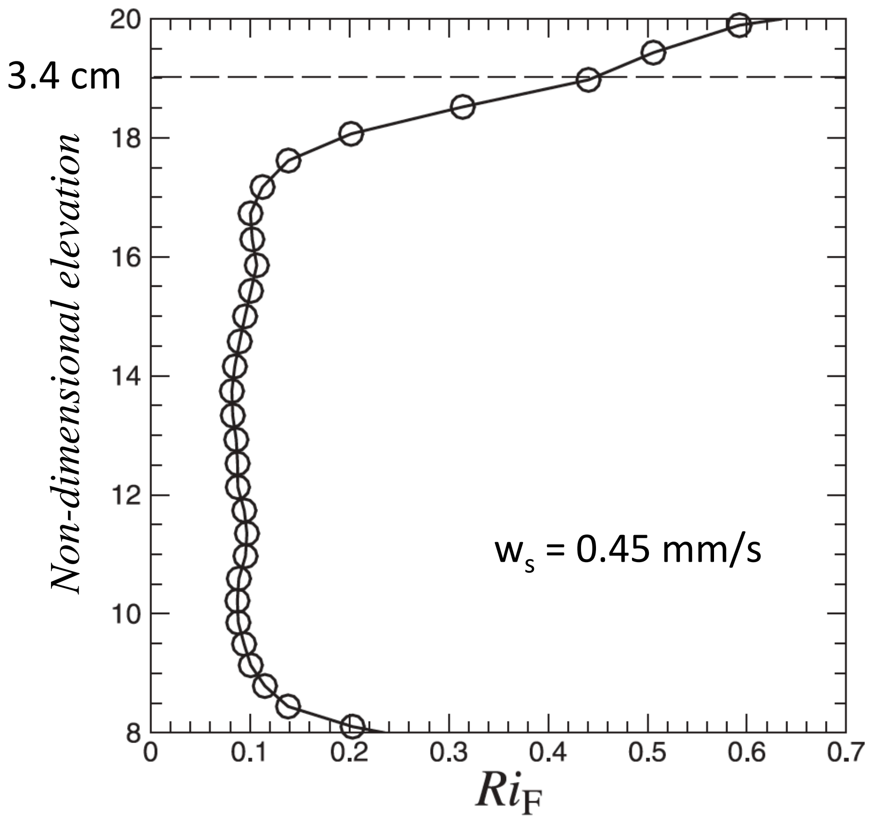


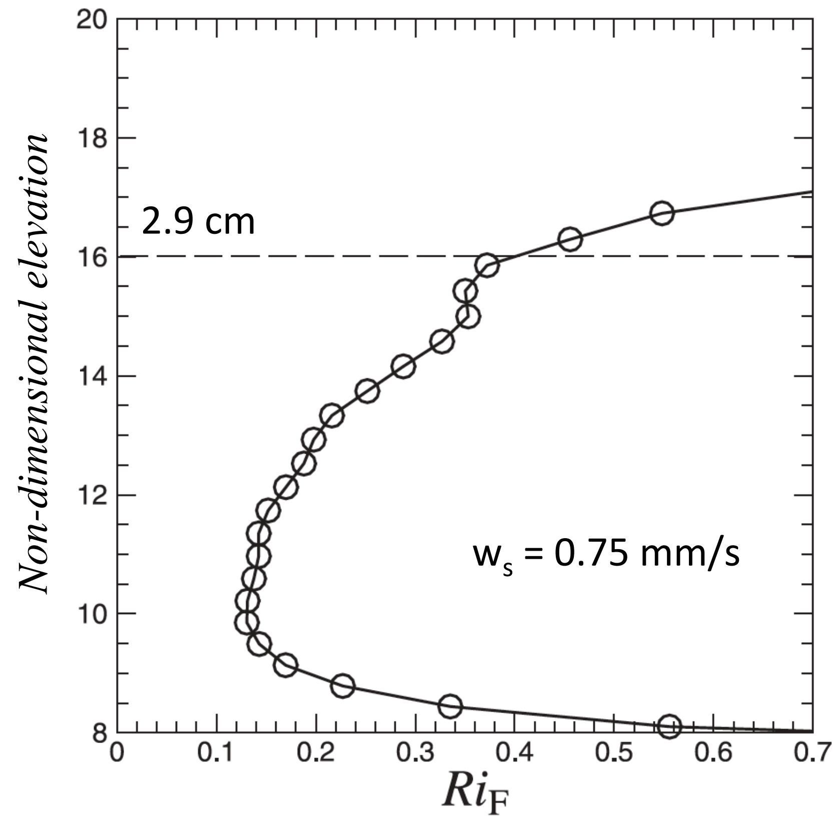
OUTLINE: 1) Ri # importance; 2) Overlap layer; 3) Under-saturation; 4) Critical Saturation; **<u>5) Over-saturation</u>**

U ~ 60 cm/s
C ~ 10 g/liter
"large eddy simulation" model

(Ozedemir, Hsu & Balachandar, in press)

Profiles of flux Richardson number at time of max free stream U





The Richardson number is of "order" critical (relatively close to 0.25) near top of suspended layer

OUTLINE: 1) Ri # importance; 2) Overlap layer; 3) Under-saturation; 4) Critical Saturation; **5) Over-saturation**

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Carl Friedrichs, Virginia Institute of Marine Science

Conclusions:

- Negative feedback favors sediment Ri ≈ Ri_{cr} in the BBL.
- Ri < 1/4 (vs. > 1/4) implies supply (vs. settling) limitation.
- Ri const. in z implies C ~ z^{-A}, with A ≈ 1 and u ~ log(z).
- If Ri↑ (vs.↓) in z, then A < 1 (vs. > 1), u concave down (vs. up).
- Ri ≈ Ri_{cr} predicts max load independent of w_s and erodibility.
- Time-scales of changes in u determine whether turbulence and suspension will catastrophically collapse via positive feedback.

Presented at University of Delaware, 9/27/11

10

1 meter

30

2 hours

100 mg/l

Time-series of suspended sediment in York River estuary (Friedrichs et al. 2000)