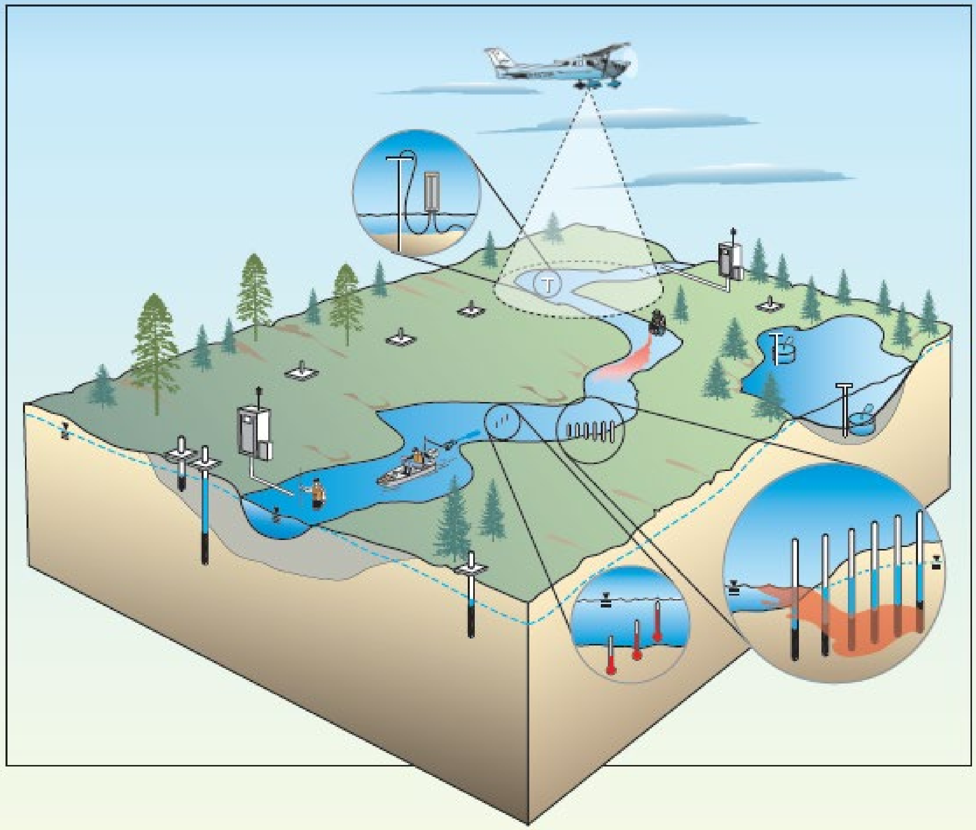
# University of Granada, 2023

## Field Techniques

Masaki Hayashi,
hayashi@ucalgary.ca
Donald Rosenberry,
rosenberry@mines.edu

Field Techniques for Estimating Water Fluxes Between Surface Water and Ground Water



Techniques and Methods 4–D2

U.S. Department of the Interior
U.S. Geological Survey

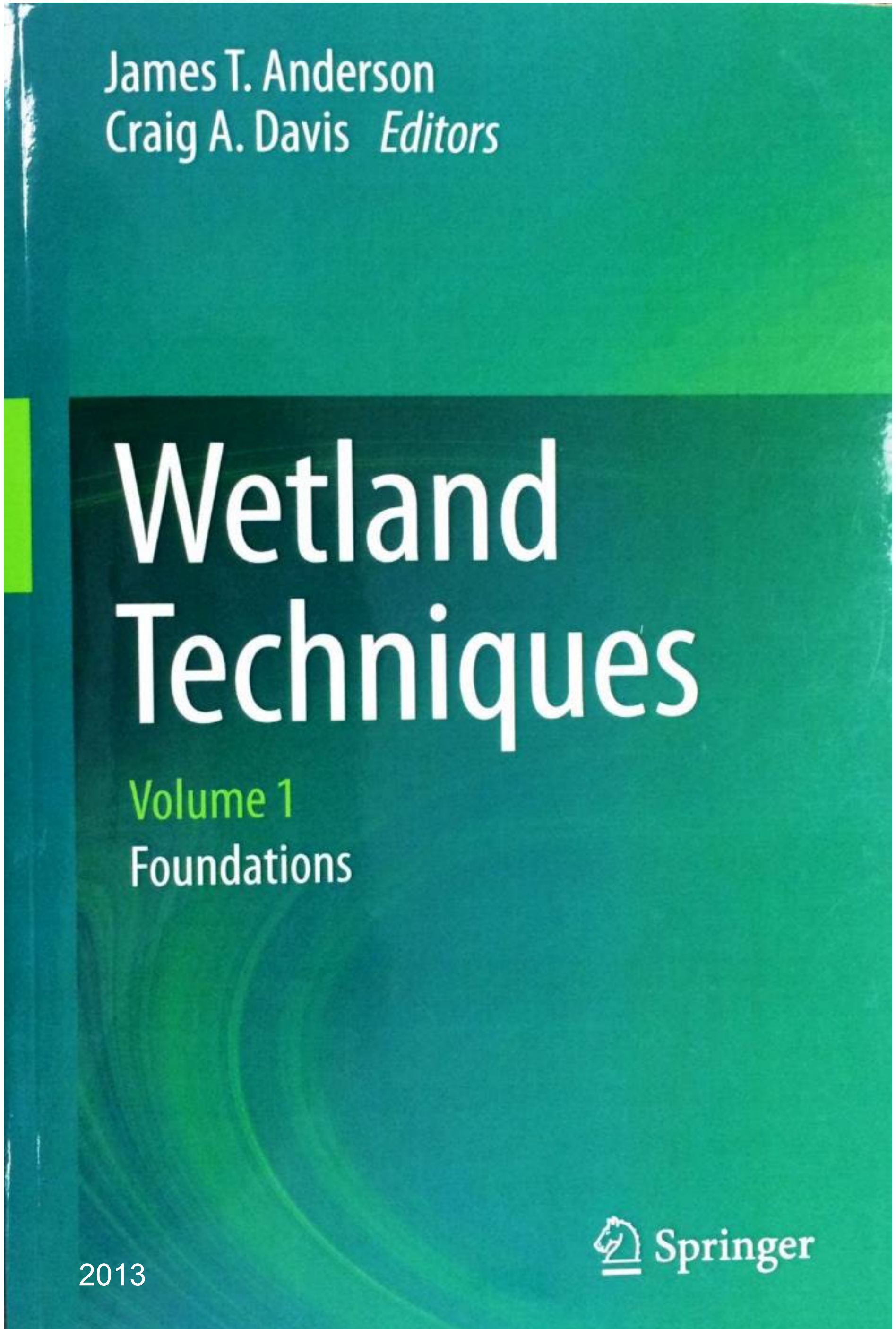
Chapter 1 – General overview

**Chapter 2 – Wells and staff gages, portable wells, seepage meters**

Chapter 3 – Tracers and methods for karst

Chapter 4 – **Heat as a tracer**

This USGS report published in 2009 presents many of the methods with which we can quantify flow between groundwater and surface water. In this lecture we will cover much of the information presented in chapter 2 of this report. The report is available for download at http://pubs.usgs.gov/tm/04d02/. We will cover information presented in Chapter 4 later, but won't talk as much about Chapter 3 as perhaps we should.




Chapter 3
**Assessing and Measuring Wetland Hydrology**

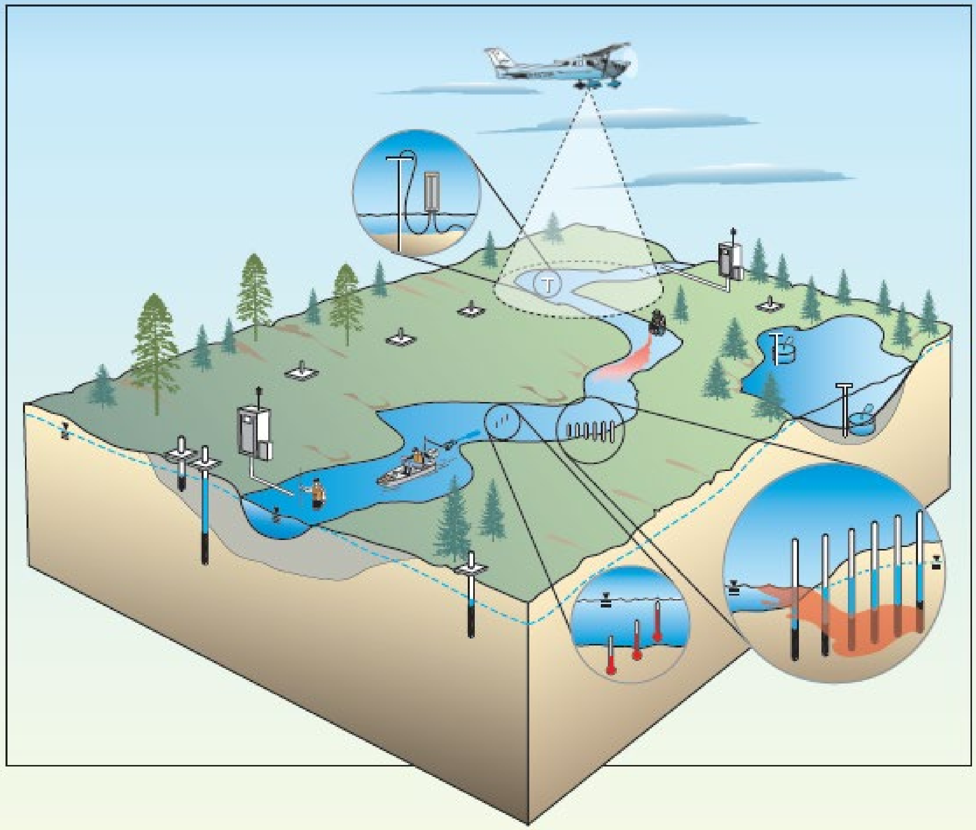
**Donald O. Rosenberry and Masaki Hayashi**



This report goes into much more detail about specific hydrological processes and how to quantify them, including step-by-step instructions for quantifying GW-SW exchange by installation of monitoring wells, surface-water stage gages, seepage meters, and temperature-based methods. It is available on request to either Masaki or Don.

USGS
science for a changing world

Field Techniques for Estimating Water Fluxes Between Surface Water and Ground Water



Techniques and Methods 4–D2

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# Choices

- Seepage run
- Measure hydraulic properties
  - Wells & SW stage
  - Portable wells
  - Seepage meters
- Aerial imagery
- Dye and tracer tests
  - Dilution gaging (Q)
  - Conservative tracer
    - Dyes, isotopes, major ions
  - Chemical mixing models
    - Major ions, isotopes
- Thermal profiling (also called vertical temperature profiling)
- Fiber-optic distributed temperature sensing (FO-DTS)
- Towed probes
- Electrical resistivity profiling (e.g., Supersting)
  - Towed surface array
  - Array planted on bed
- Biological indicators
- Water and chemical budgets (GW as residual)

We will discuss most of the items on this list. Several other methods also are presented in this course. Methods shown in green are presented in a subsequent talk because use of temperature to quantify GW-SW exchange are now so commonly used.

The seepage run consists of measuring streamflow at the upstream and downstream ends of a stream reach, and then assuming that the difference is due to net groundwater discharge to the stream or loss of streamwater to groundwater. This method represents groundwater-surface-water exchange on a watershed or sub-watershed scale.

# Seepage run

- Lemhi River, Idaho
- Very detailed segmenting of the river
- Two seepage runs, one during irrigation and the other after irrigation

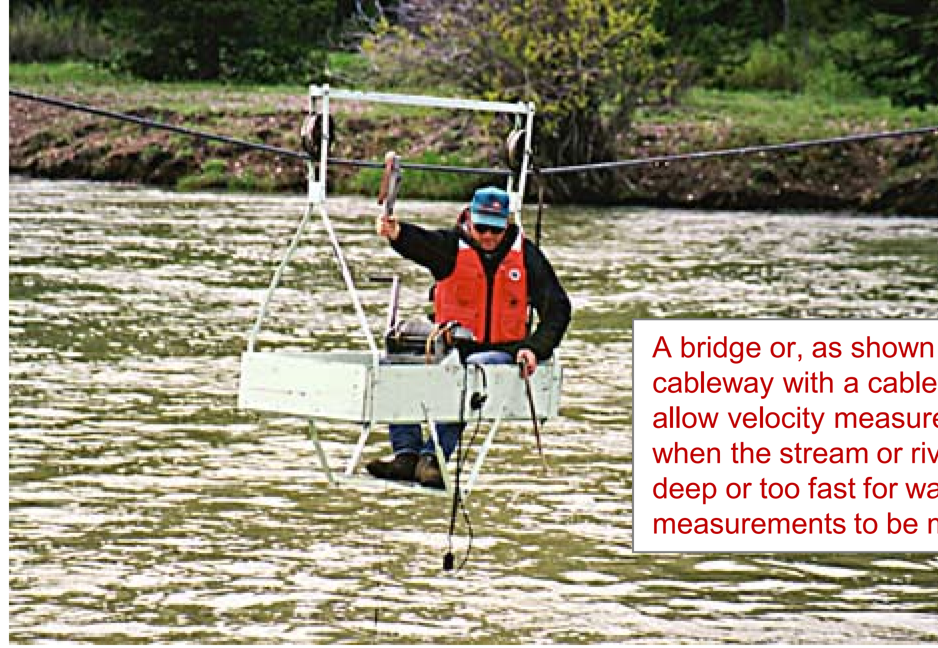
Donato, 1998, WRIR 98–4185


Stream discharge is measured at each black dot on the map. Values for net GW exchange are determined for each numbered stream reach.

This method is sometimes called differential gaging.

A few methods for measuring velocity

A wading rod with attached flowmeter ("old school")



A bridge or, as shown here, a cableway with a cable car, allow velocity measurements when the stream or river is too deep or too fast for wading measurements to be made.

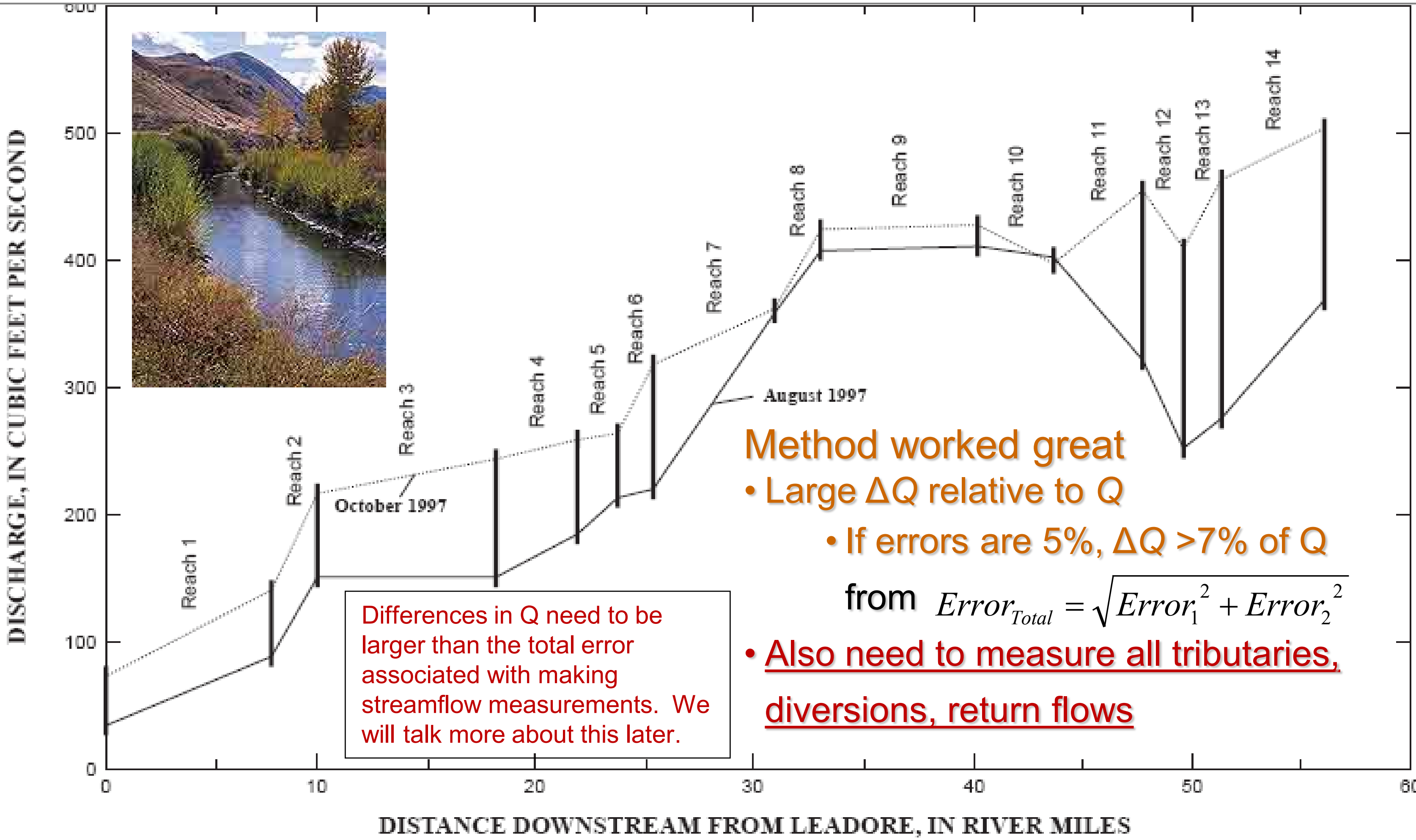
This is a portable flume that can be installed to measure flow when discharge is very small.

A Price AA or Pygmy meter (shown here) has long been the standard. Most users have switched to acoustic doppler velocimeters (ADVs) instead.


ADV


This is one of a variety of acoustic devices that can provide flow in 3 dimensions, determine stream depth, and calculate total *Q* (note the on-board GPS receiver that determines the precise location of the device.  In this figure, a rope and pulley are used to pull the flowmeter across the stream section.

Flow increased from reach 1 to reach 8, indicating that groundwater was discharging to the upper reaches of the stream. There was very little net groundwater discharge along reaches 9 and 10, and a substantial loss of streamwater to groundwater occurred along reaches 11 and 12 during the August measurements. By comparing seepage runs during August and October (assuming there was no irrigation during October) we can see the effect of pumping groundwater on flow in the stream.




Differences in Q need to be larger than the total error associated with making streamflow measurements. We will talk more about this later.

## Method worked great

- Large ΔQ relative to Q
  - If errors are 5%, ΔQ >7% of Q

from $Error_{Total} = \sqrt{Error_1^{\,2} + Error_2^{\,2}}$

- <u>Also need to measure all tributaries, diversions, return flows</u>

Figure 6. Measured instantaneous discharge in the Lemhi River, east-central Idaho, August and October 1997. (Locations of reaches shown in figure 4)

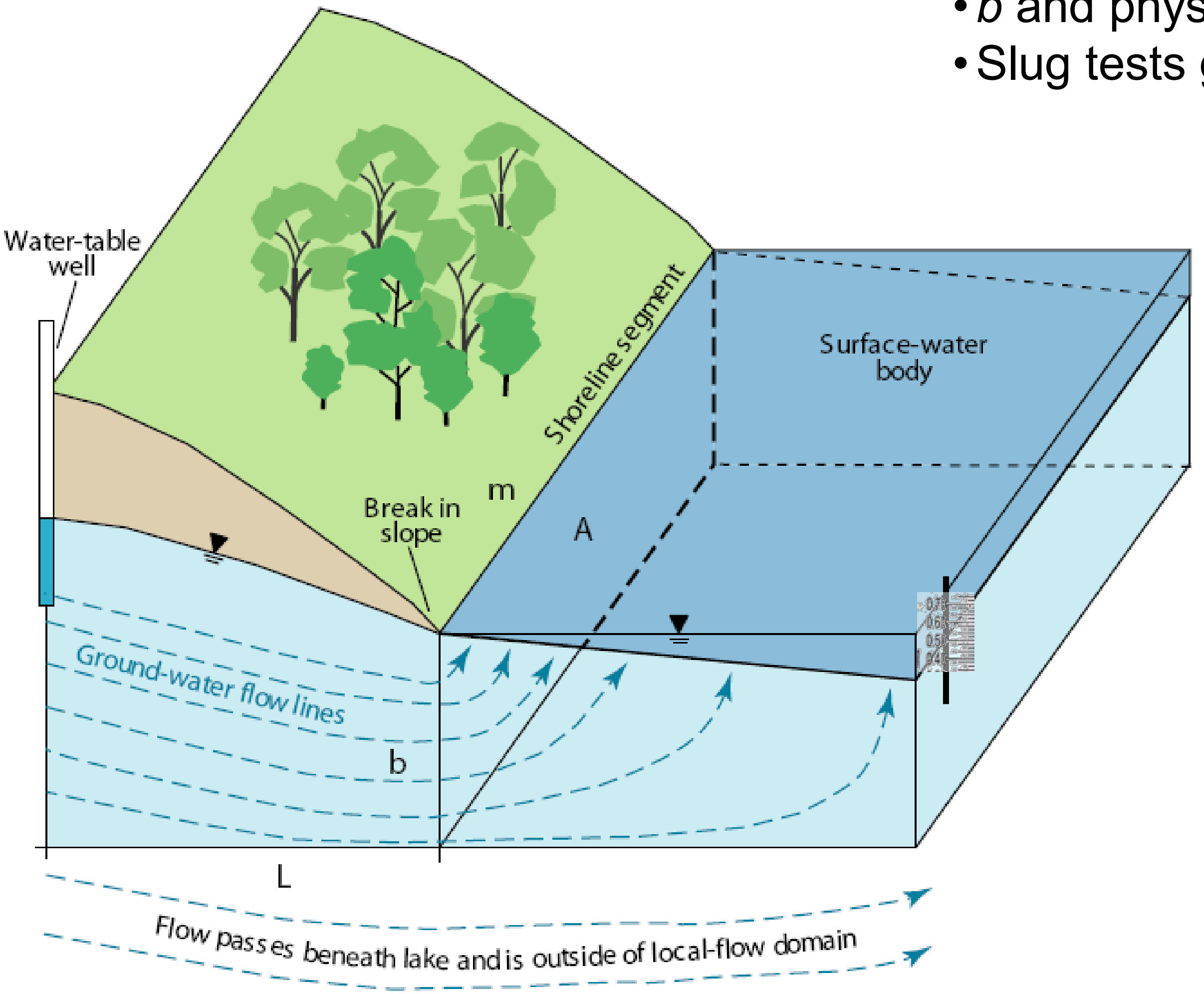
Donato, 1998, WRIR 98–4185

# Direct measurement of hydraulic properties

$$Q = KA\frac{h_1 - h_2}{L}$$

Wells and staff gage

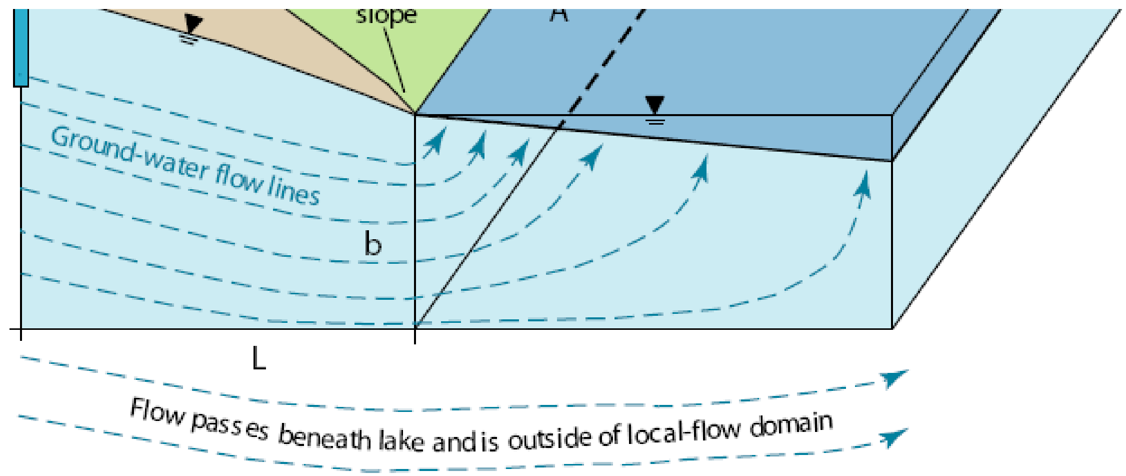
- Gives you $\Delta h$ and $L$
- Knowledge of geology gives you $b$
- $b$ and physiography gives you $A$
- Slug tests give you $K$




If we measure or estimate *K*, *A*, and the hydraulic gradient, we can calculate *Q* using Darcy's Law for specific segments of a surface water body. This message includes a few assumptions and challenges that we will discuss. It is one of the most commonly used methods for determining *Q*.

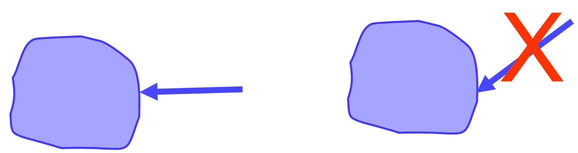
# Wells and staff gage – Assumptions

1. All water that exchanges with a surface-water body passes horizontally through a vertical plane positioned at the shoreline that extends to a finite depth (*b*) beneath the surface of the surface-water body. At depths greater than *b*, ground water flows beneath the surface-water body and does not exchange with the surface-water body.

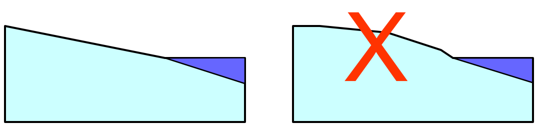



The extent of the vertical plane ***(b)*** is rarely known and we often must estimate this term or assign an uncertainty range. Violation of assumptions 2 and 3 often result in small errors. We nearly always assume that the aquifer is homogeneous and isotropic, although that rarely is true. Errors from violating these assumptions usually are small compared to the uncertainty in determining *K*.

2. The direction of water flow is perpendicular to the shoreline.

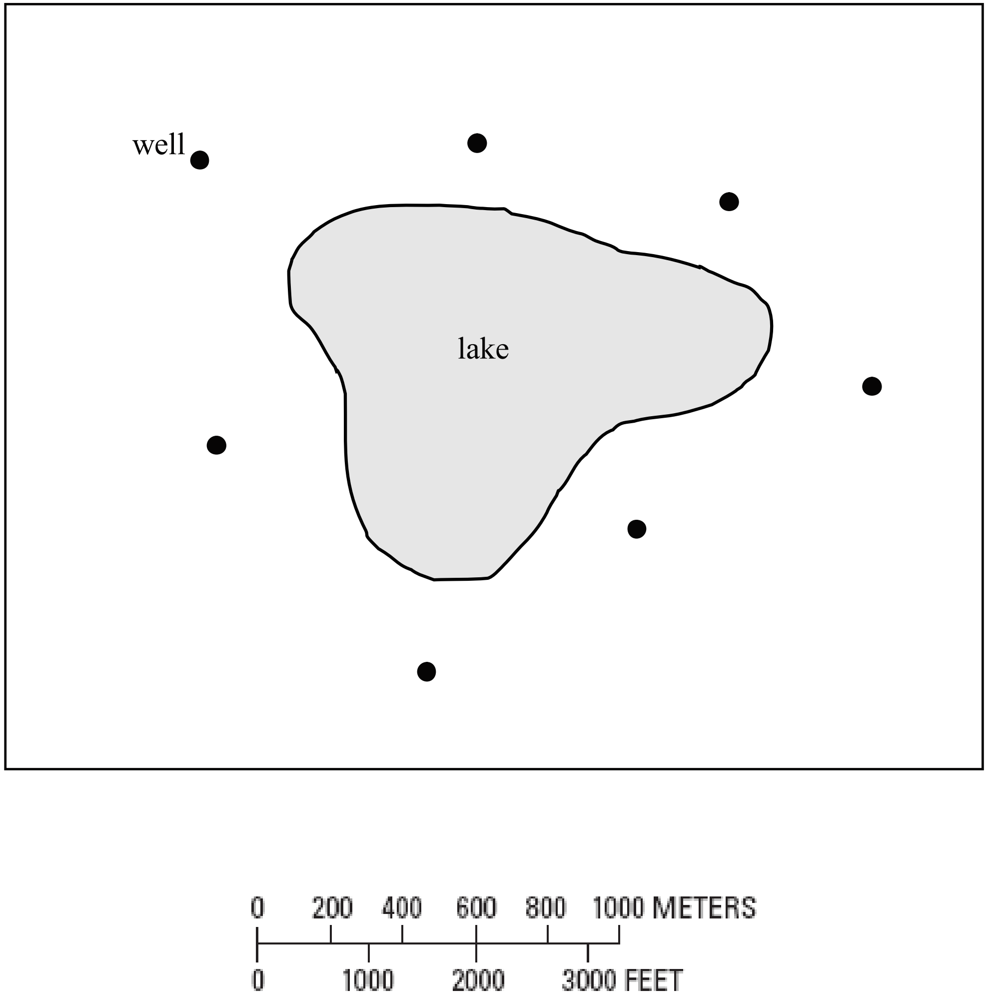



3. The gradient (water-table slope) between the well and the surface-water body is uniform.



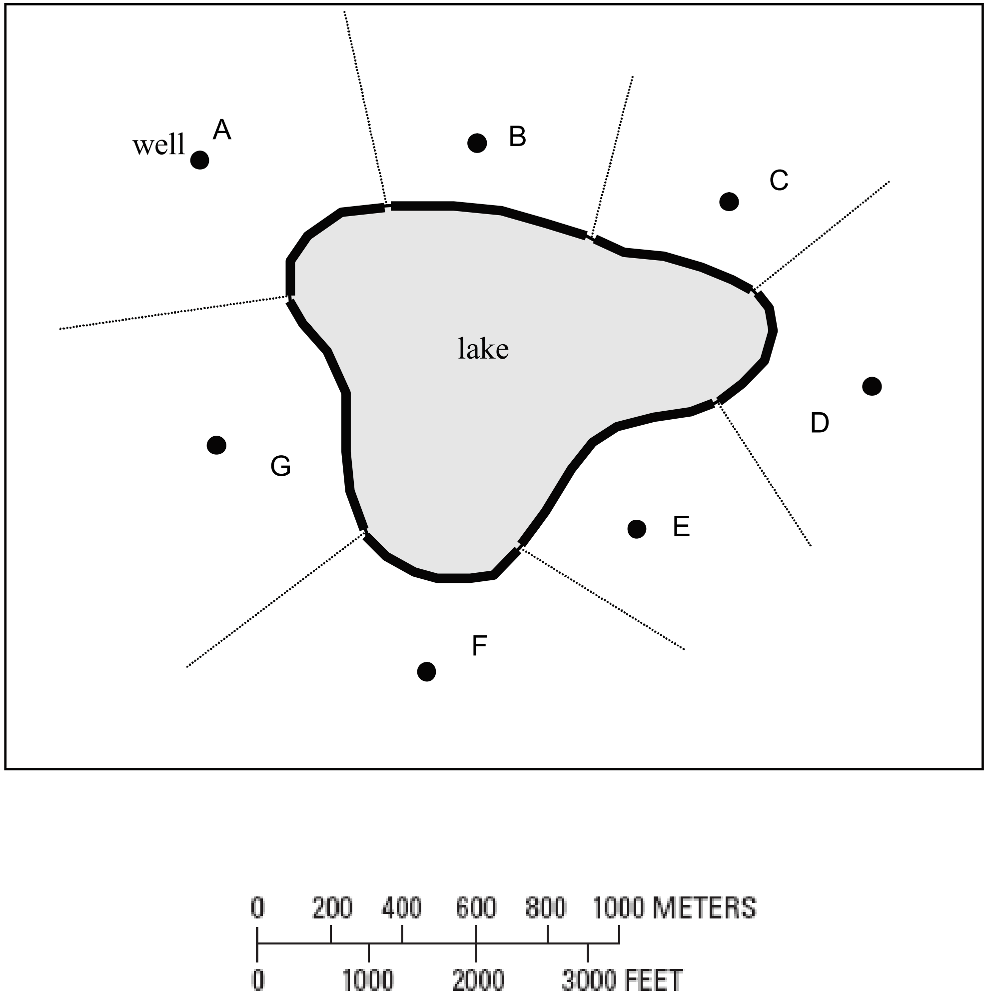

4. The aquifer is homogeneous and isotropic within the segment.

# Segmented-lakeshore approach




In this setting we have 7 monitoring wells distributed approximately evenly around a lake. How fortunate we are! Our plan will be to make water-level measurements in each of the wells, measure the water level of the lake, survey all measurements to a common datum, and then determine flow direction and hydraulic gradients within the aquifer surrounding the lake, as well as between the aquifer and the lake. Pretty simple, right?

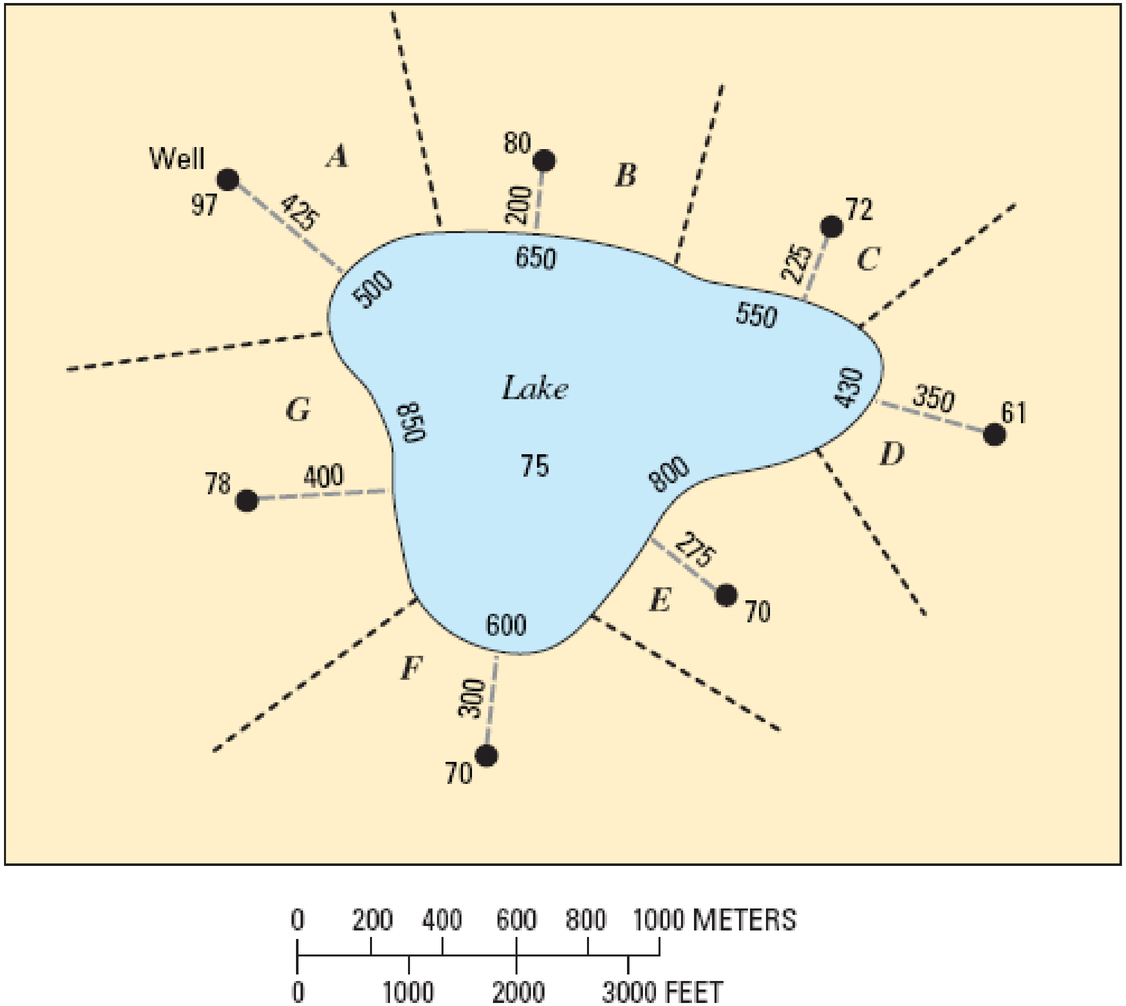
# Segmented-lakeshore approach

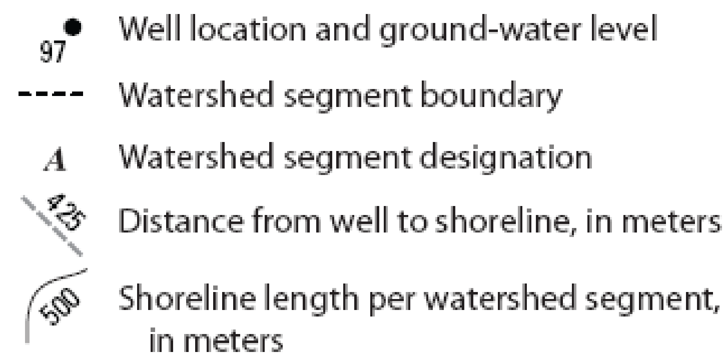



In this method, called the segmented-lakeshore method, we assign a segment of the shoreline of the lake to each monitoring well. We then draw lines that intersect the lake shoreline and extend approximately perpendicularly away from the shoreline at the end of each segment. We will assign a uniform hydraulic gradient for each of the areas bounded by the lines that we have just drawn. Does this seem reasonable?

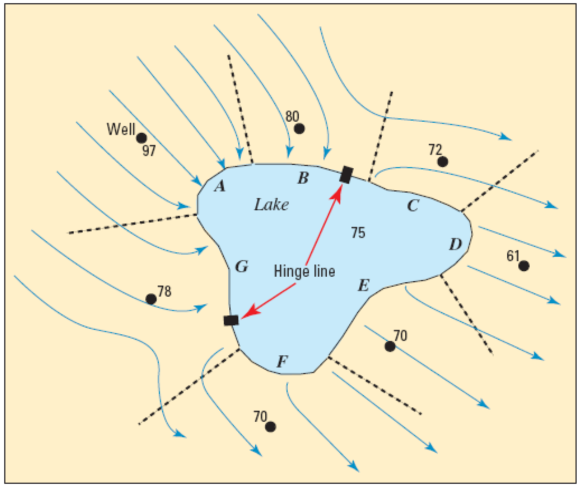
# Segmented-lakeshore approach

$$Q = KA\frac{h_{GW} - h_{LK}}{L}$$




Next, we measure the shoreline lengths (used to determine *A*), the distance from the well to the shoreline (*L*), and we determine $h_1$-$h_2$ between each well and the lake by making depth-to-water measurements and measuring the lake stage. We then make these observations relative to one another by surveying wells and the lake staff gage. The values of the lake (75) and the wells (e.g., 97 or 80) are in meters above our common datum.




# Segmented-lakeshore approach

Total $Q_{in}$ = 29,104 m³/d

Total $Q_{out}$ = 29,447 m³/d

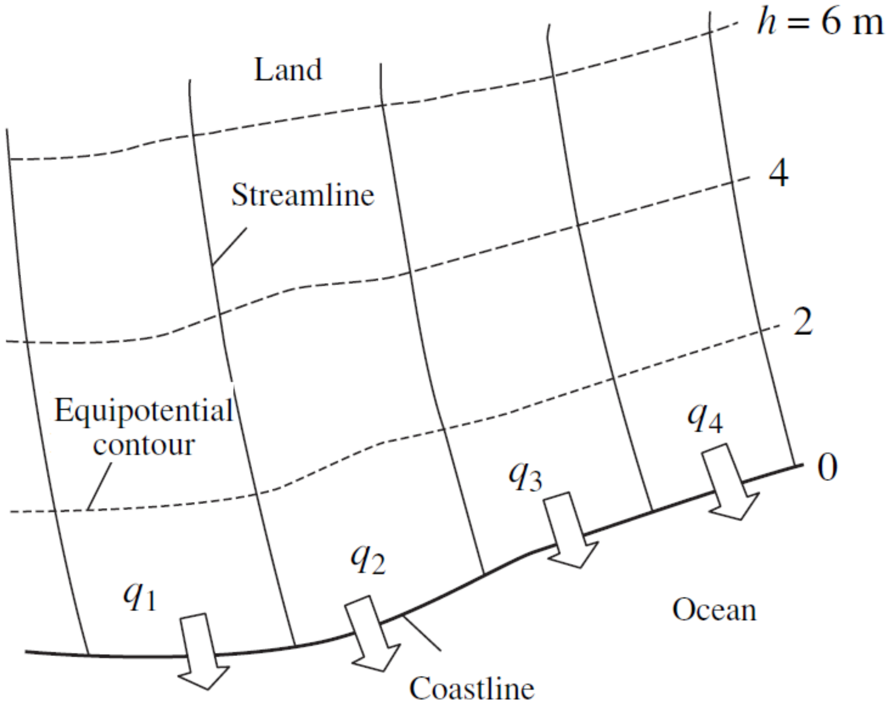
$$Q = KA\frac{h_{GW} - h_{LK}}{L}$$

You can see we have assumed that *b* is 20 m and that *K* is uniform everywhere and is 30 m/d. We have entered the values into this spreadsheet and have calculated *Q* for each shoreline segment. Although $Q_{in}$ and $Q_{out}$ don't have to match (other components of the water balance could make up the difference), they are nearly equal in this example.

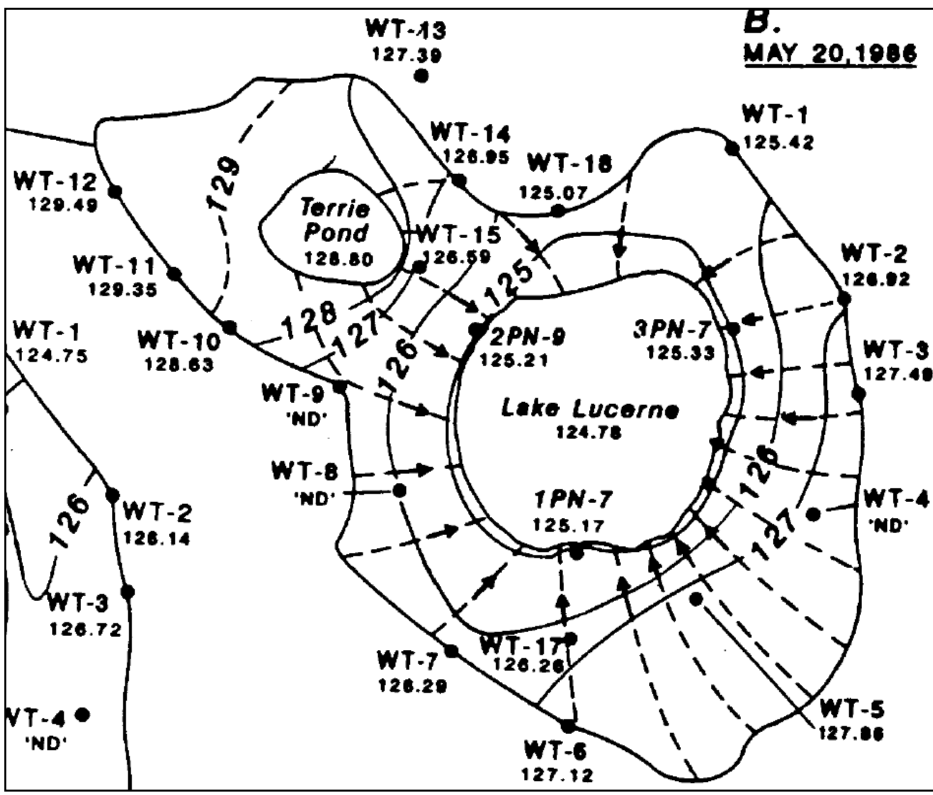
| Watershed segment | Horizontal hydraulic conductivity (*K*), m/d | Effective thickness of the aquifer (b), m | Hydraulic head in well minus surface-water stage ($h_{GW}$–$h_{LK}$), m | Distance from the well to the shoreline (*L*), m | Length of shoreline segment (*m*), m | Water flow (*Q*), m³/d |
|---|---|---|---|---|---|---|
| A | 30 | 20 | 22 | 425 | 500 | 15,529 |
| B | 30 | 20 | 5 | 200 | 650 | 9,750 |
| C | 30 | 20 | –3 | 225 | 550 | –4,400 |
| D | 30 | 20 | –14 | 350 | 430 | –10,320 |
| E | 30 | 20 | –5 | 275 | 800 | –8,727 |
| F | 30 | 20 | –5 | 300 | 600 | –6,000 |
| G | 30 | 20 | 3 | 400 | 850 | 3,825 |

# Flow-net approach

The flow-net approach for quantifying flows to or from a surface-water body has been around for many decades. It is not very commonly used but is a good "ballpark" method for generating quick estimates. If you are interested in trying this, two good examples of the use of this method are cited below. The method also uses Darcy's Law to calculate flow, but the equation looks slightly different relative to what we are used to seeing.




Loaiciga, H. A., and I. S. Zektser (2003), Estimation of submarine groundwater discharge, *Water Resources*, *30*(5), 517-524.

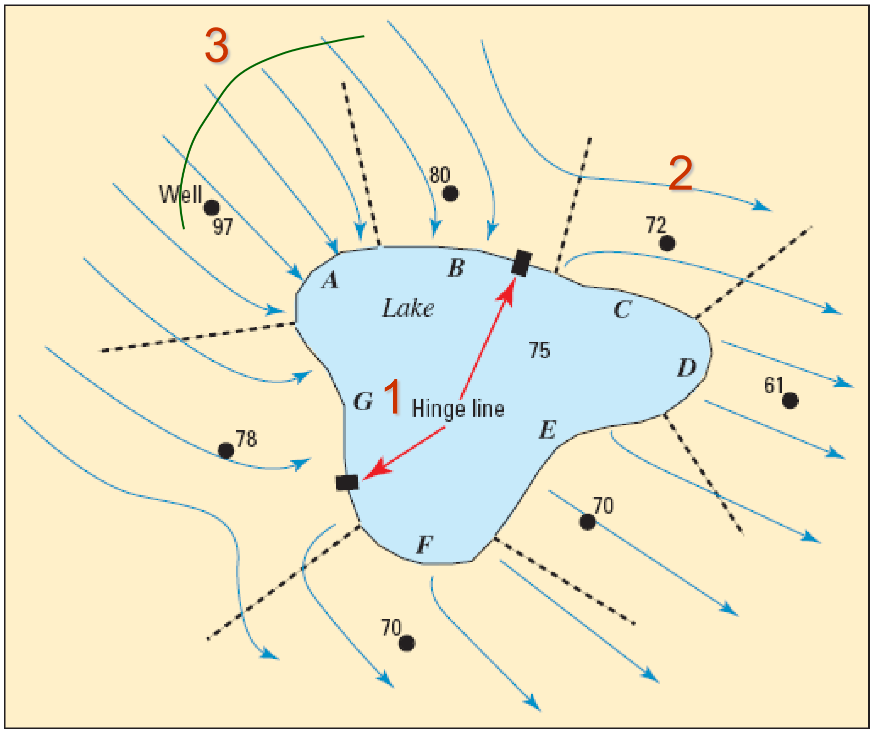



Lee, T. M., and A. Swancar (1997), Influence of evaporation, ground water, and uncertainty in the hydrologic budget of Lake Lucerne, a seepage lake in Polk County, Florida, *Water-Supply Paper*, 61 pp, U.S. Geological Survey.

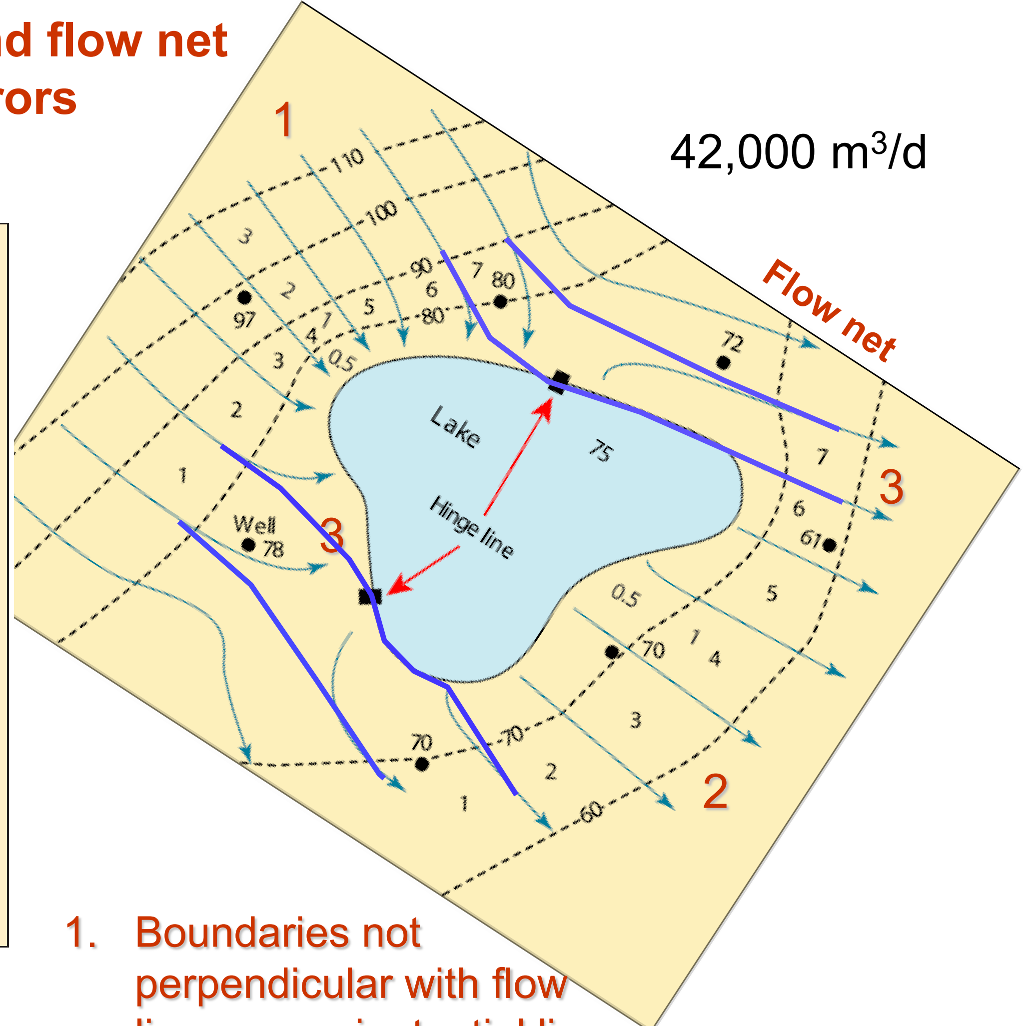
# Comparison of segmented Darcy and flow net results, and associated potential errors

29,000 $m^3/d$

**Segmented Darcy**




42,000 $m^3/d$

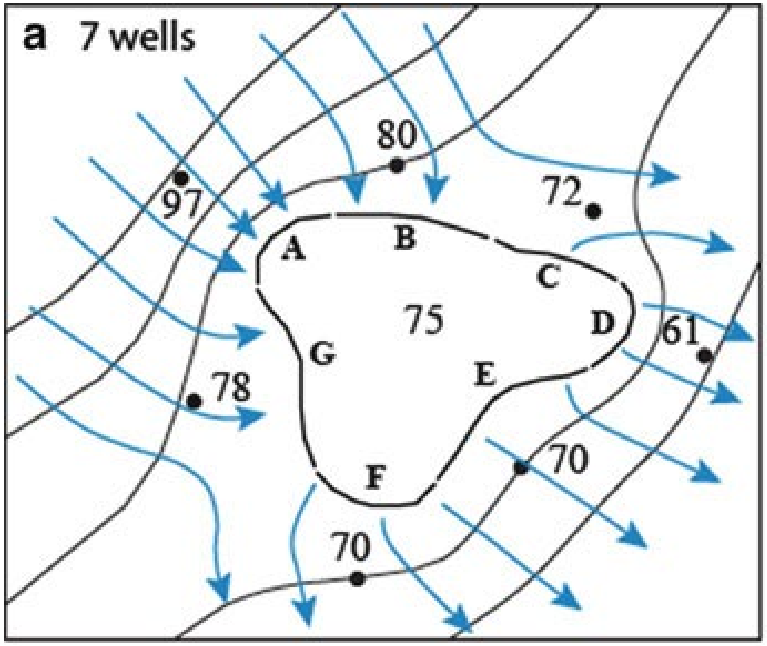



1. Segment boundaries don't match hinge lines
2. Flow between well and lake is not perpendicular to shoreline
3. Head at well is not representative of entire shoreline segment

1. Boundaries not perpendicular with flow lines or equipotential lines
2. Discretization too coarse to represent flow complexity (only 1.5 "head drops")
3. Incorrect interpretation of streamtubes intersecting the lake

**If tubes 1 and 7 are ignored, Q = 30,000 $m^3/d$**

# No matter the method, how many wells do we need?

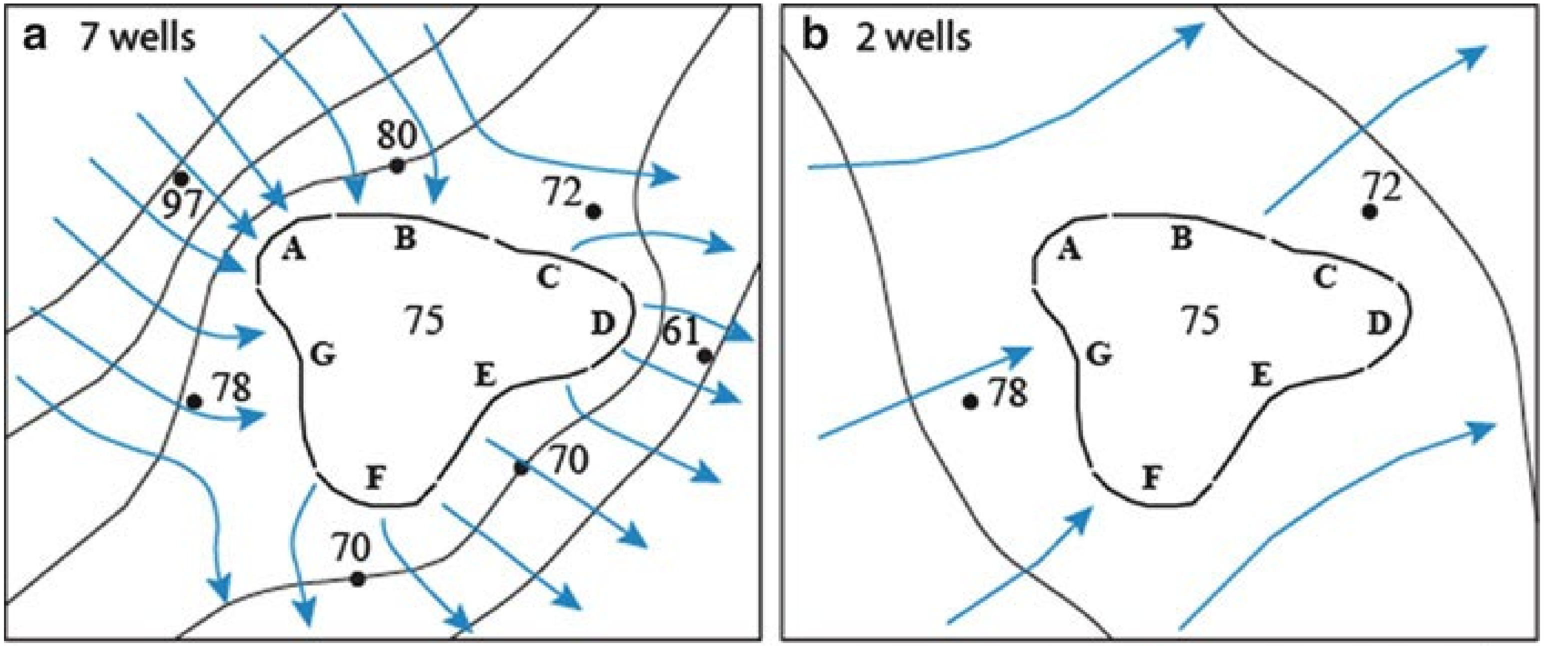



Seven wells surrounding a small lake is fantastic; we can be confident in our interpretation of flow with this number of data points. But this richness of data is not very common. I've been asked numerous times, "What is the minimum number of wells that we need?"

Rosenberry and Hayashi, 2013, Wetland Hydrology book chapter, Springer

Lots of how-to hands-on information about water-budget terms.

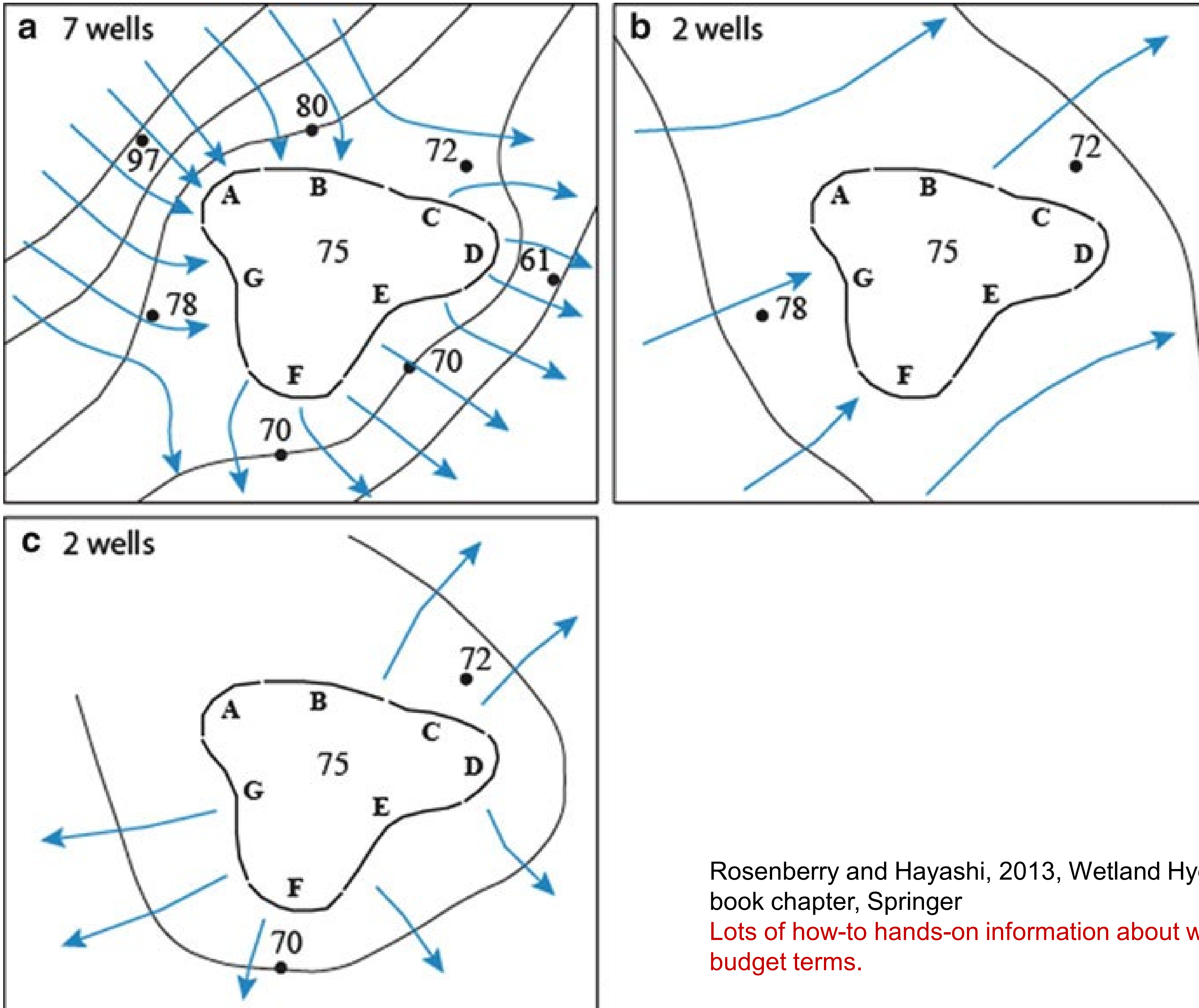
# Two wells give us a different flow direction




Rosenberry and Hayashi, 2013, Wetland Hydrology book chapter, Springer

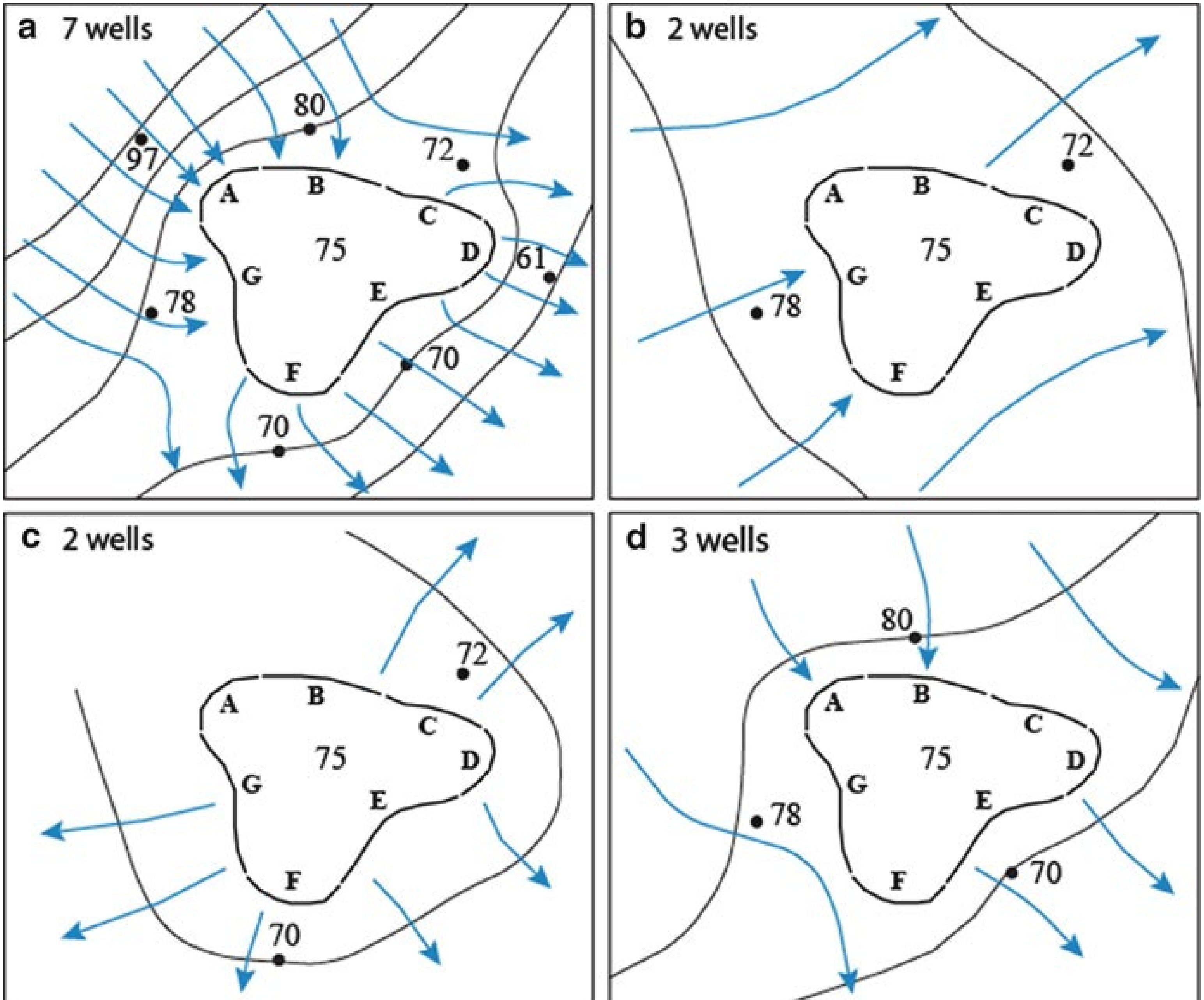
Lots of how-to hands-on information about water-budget terms.

# And a different combination of two wells shows a recharge lake




Rosenberry and Hayashi, 2013, Wetland Hydrology book chapter, Springer
Lots of how-to hands-on information about water-budget terms.

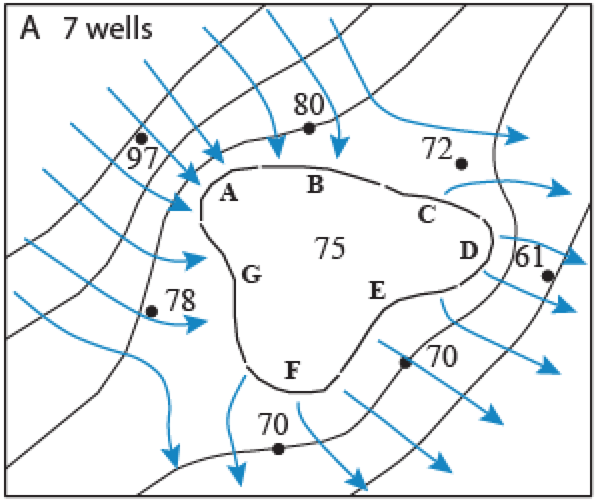
# Three wells comes pretty close

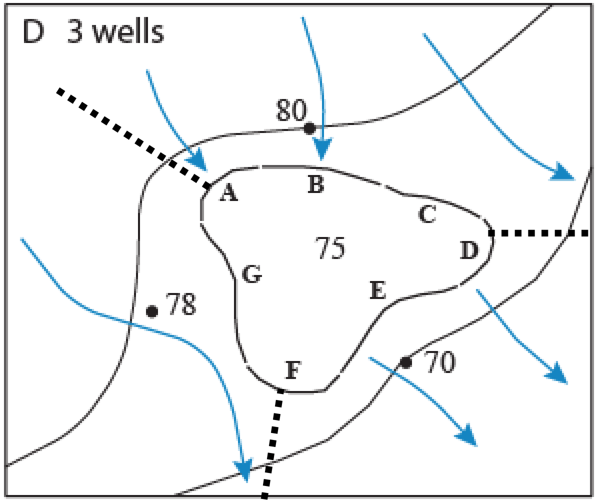


# Interpretation depends on the available data

In = 29,100

Out = -29,500




In = 25,500

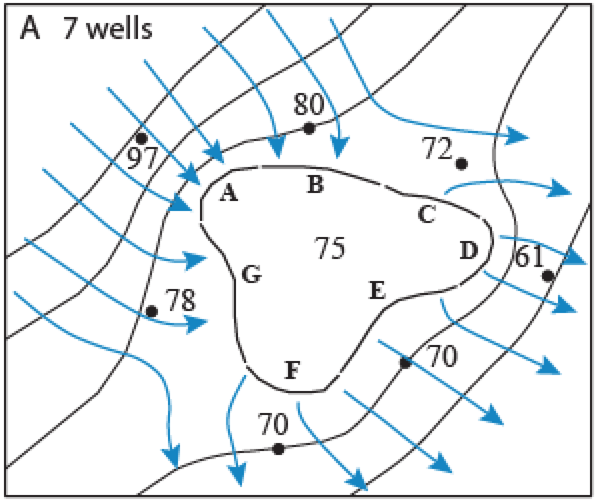
Out = -20,000

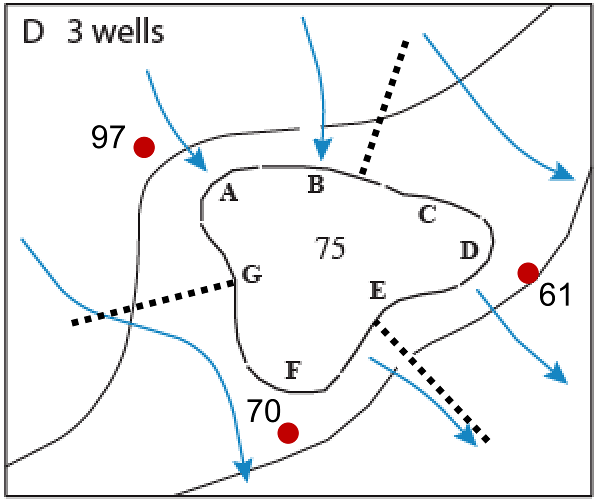
These three wells give us flows to and from the lake that match pretty closely the values when we have 7 wells.

# Interpretation depends on the available data

In = 29,100

Out = -29,500




In = 62,000

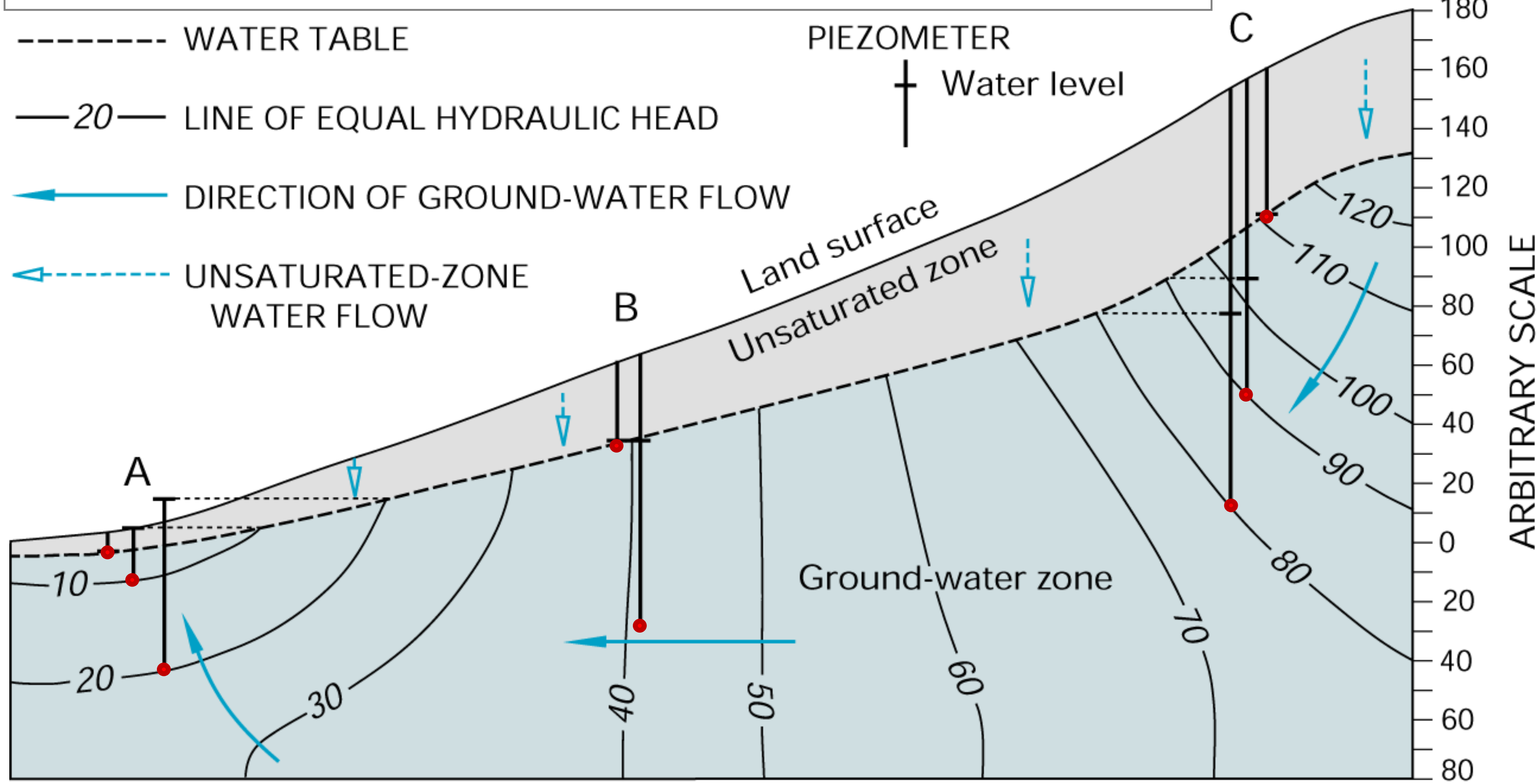
Out = -43,000

But the total flows to and from the lake change a lot if we use a different combination of three wells.

# Wells can indicate more than just horizontal flow

GW flow in cross section – piezometers indicate potential for flow

Do you remember this slide from Day 1? Piezometers are a special type of well than can give us very useful information. In combination, they can give us vertical hydraulic gradient.




**GW discharge -- The deeper the well the higher the head**

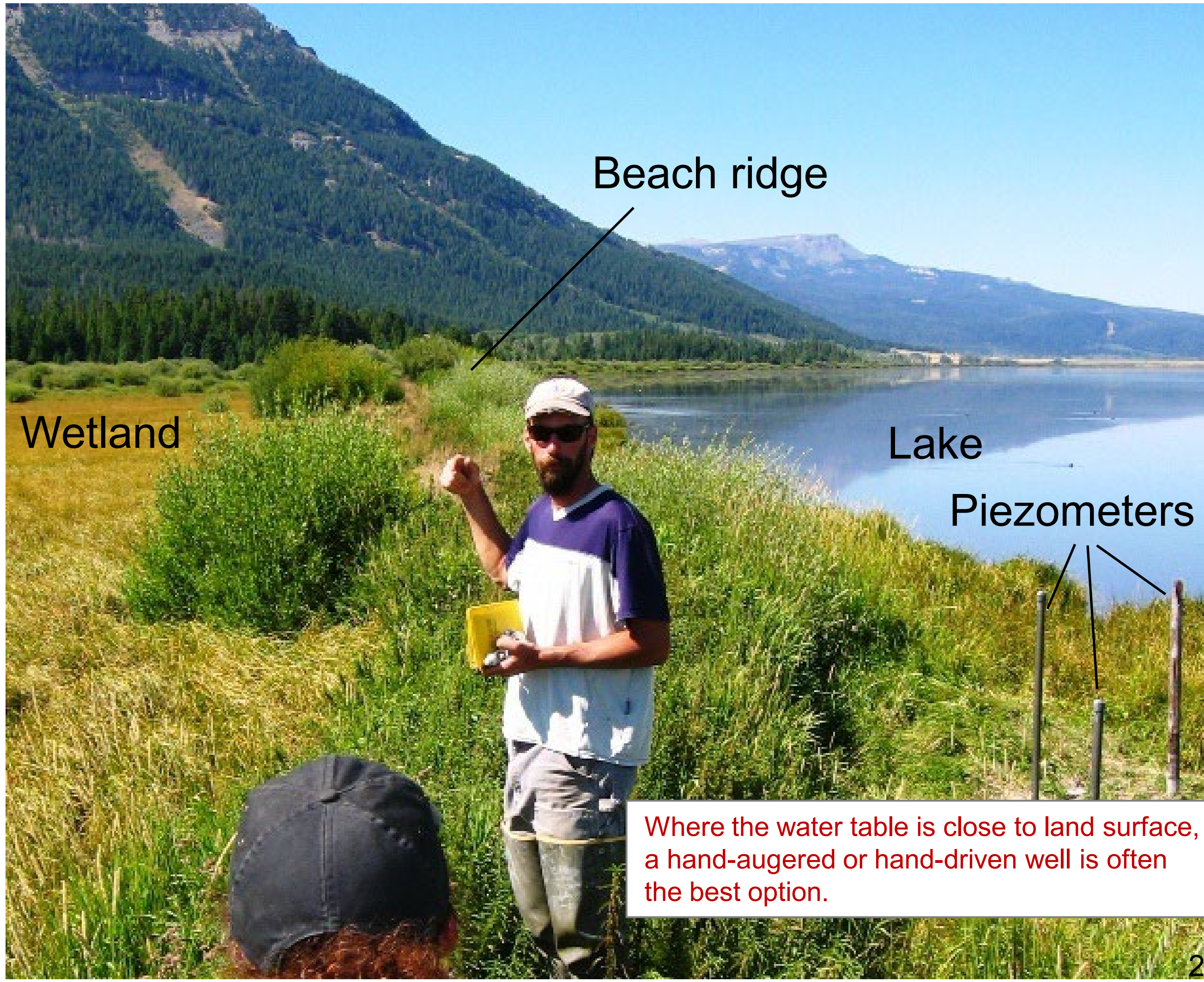
**All wells have the same head no matter the screen depth**

**GW recharge -- The deeper the well the lower the head**

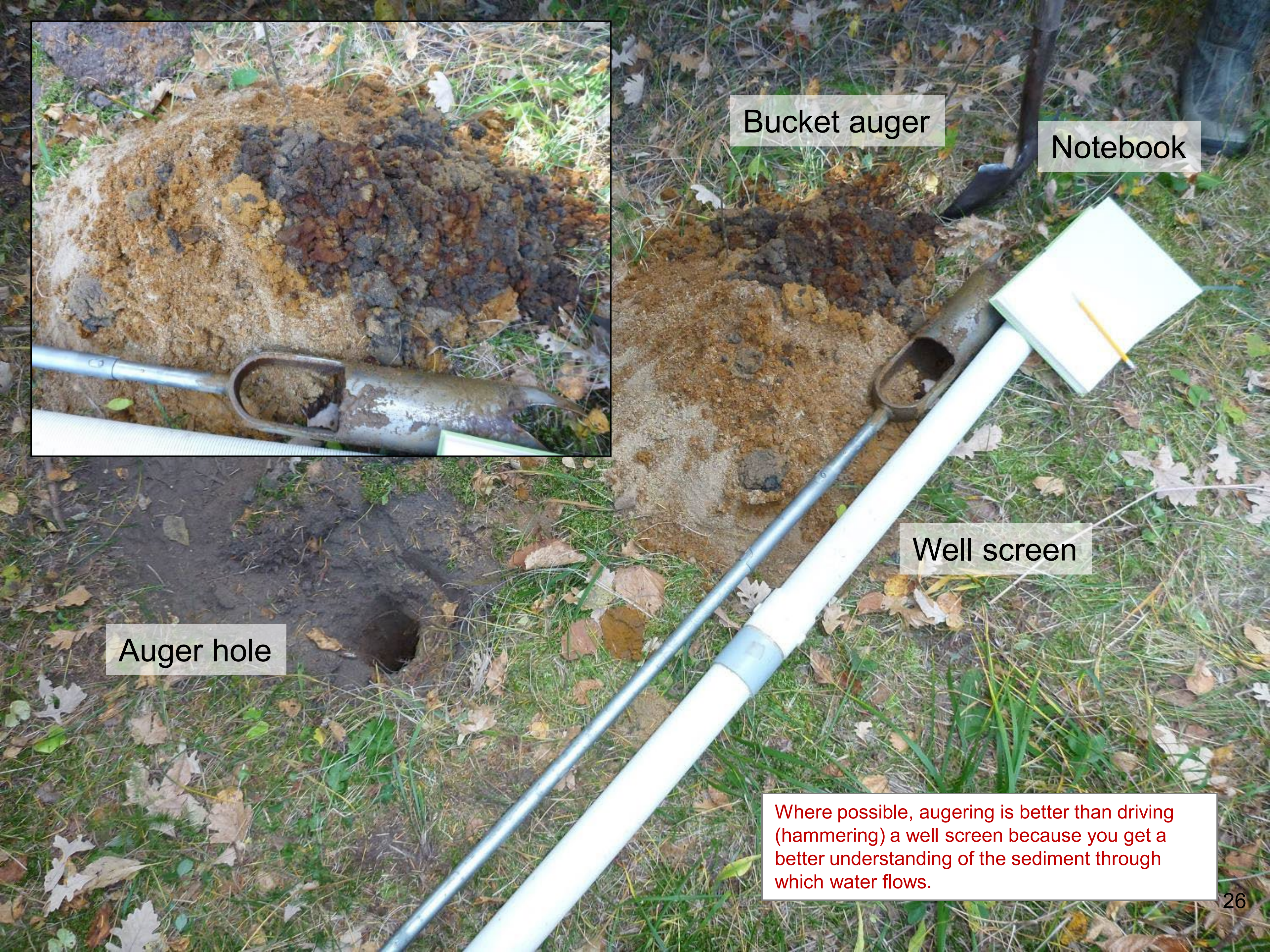
# Installing wells usually implies big and expensive equipment


Installing a piezometer in this way requires a pretty large drill rig for deeper installations. But we don't have to use large, expensive equipment if the piezometer is not very deep.

# But it can be difficult to get a large drill rig close to shore in many settings

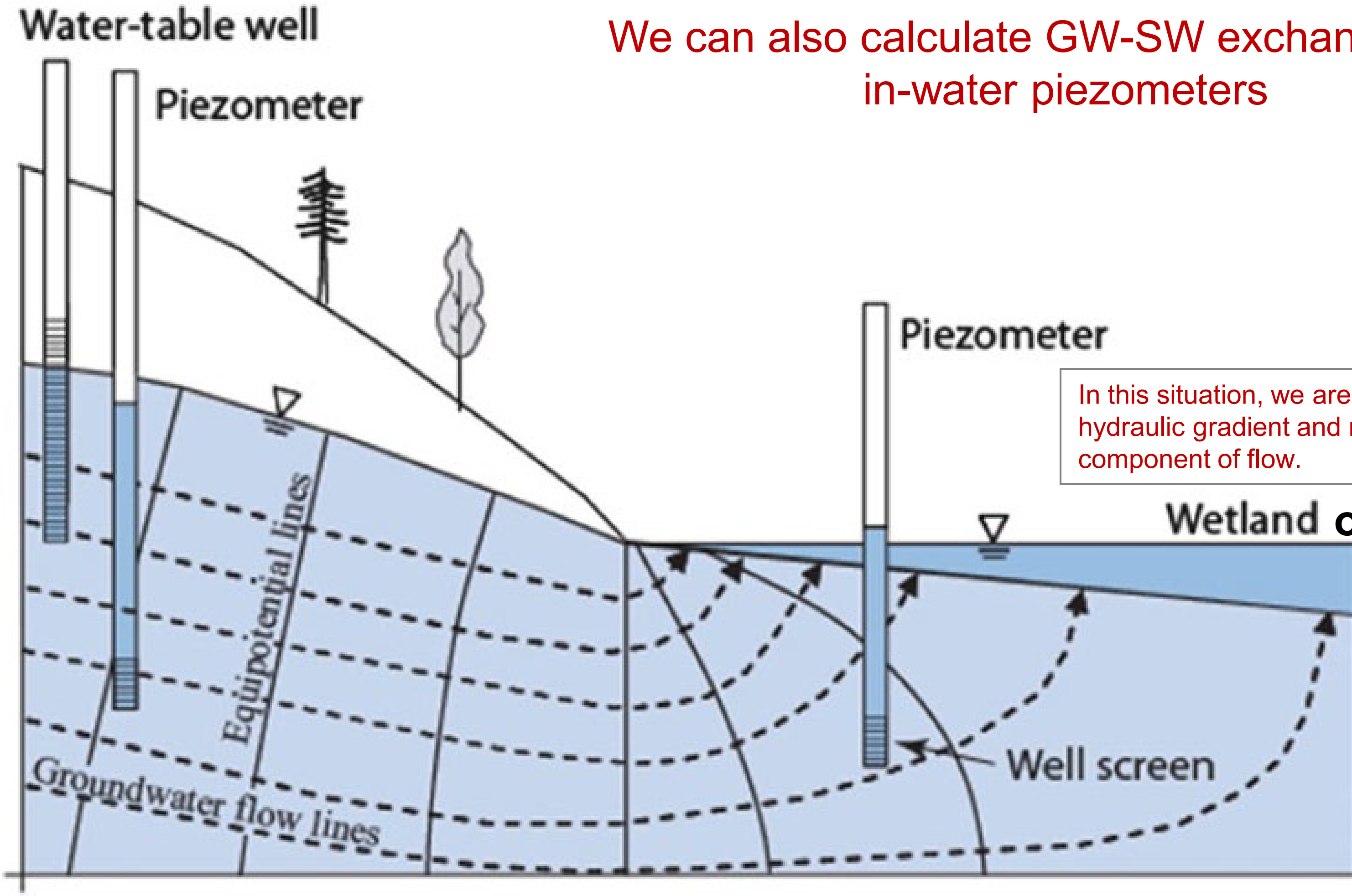


For installations less than 6 m or so below land surface, the piezometer can be installed with manual methods.  Here we are showing the use of a manual soil auger.  Different augers are available for a range of soil types.




Where possible, augering is better than driving (hammering) a well screen because you get a better understanding of the sediment through which water flows.

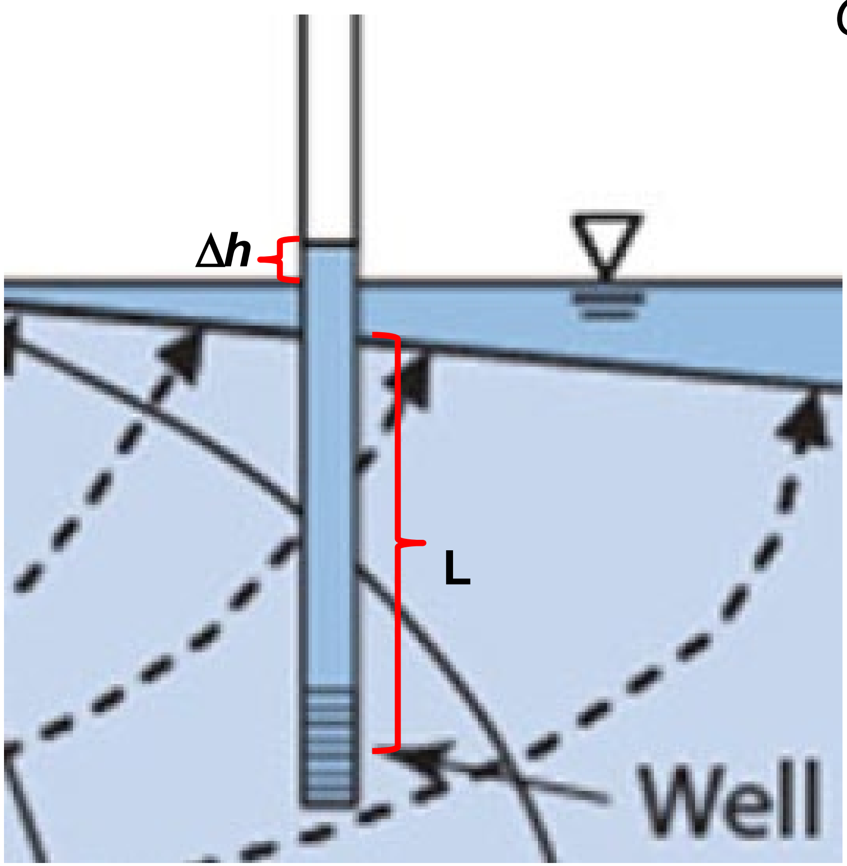
We can also calculate GW-SW exchange using in-water piezometers




In this situation, we are measuring a vertical hydraulic gradient and measuring the vertical component of flow.

From Rosenberry and Hayashi, 2013, Chapter 3 – Wetland hydrology

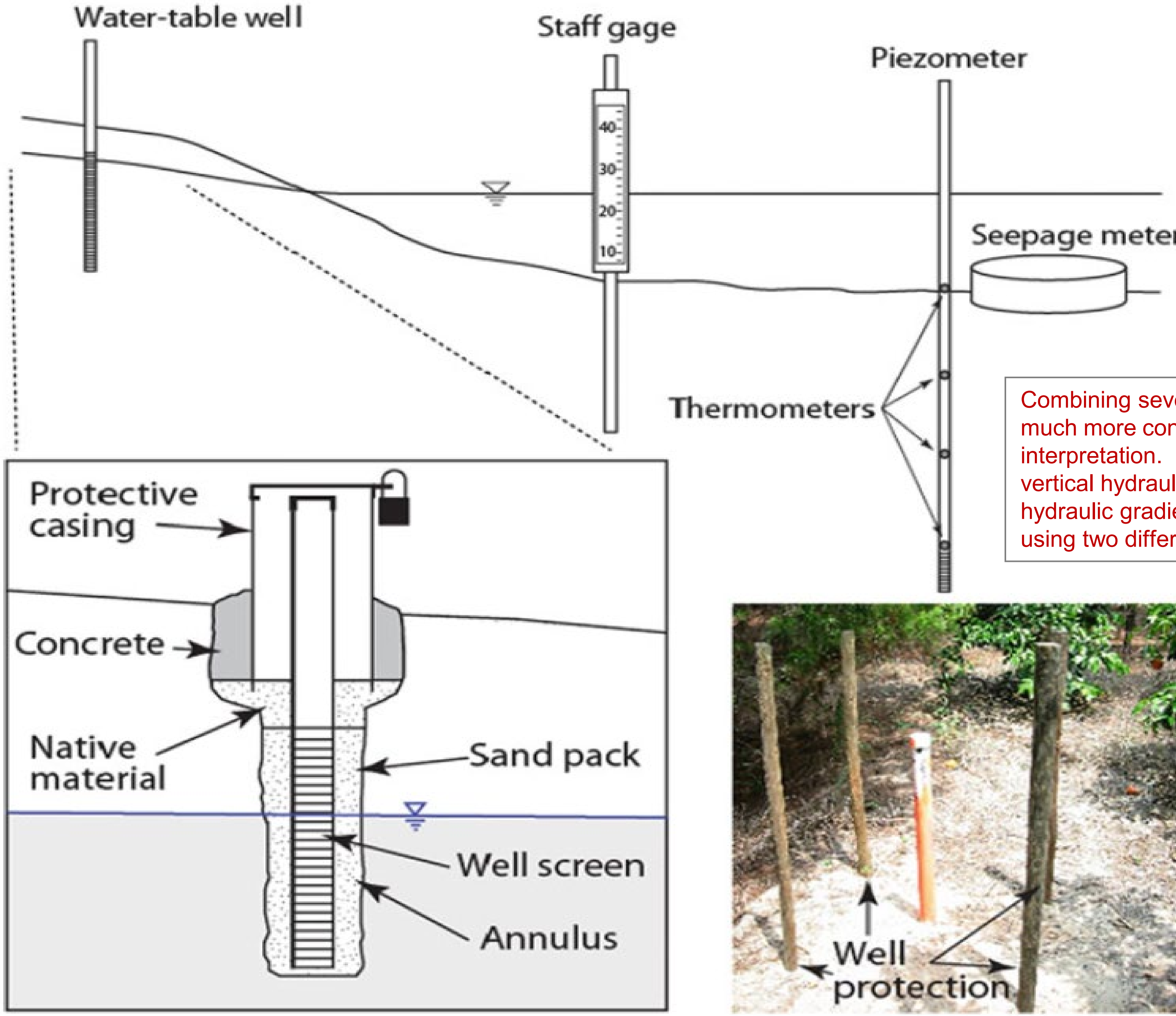
**Fig. 3.24** Depiction of the two types of monitoring wells commonly installed in wetland settings. The water level in the piezometer on the left is lower than the water level in the water-table well, indicating that a downward gradient exists in the aquifer at that point. The piezometer to the right positioned in the wetland has a water level higher than the wetland surface, indicating an upward gradient exists in the aquifer beneath the wetland




$$Q = \frac{KA\Delta h}{L}$$

$$q = \frac{K\Delta h}{L}$$

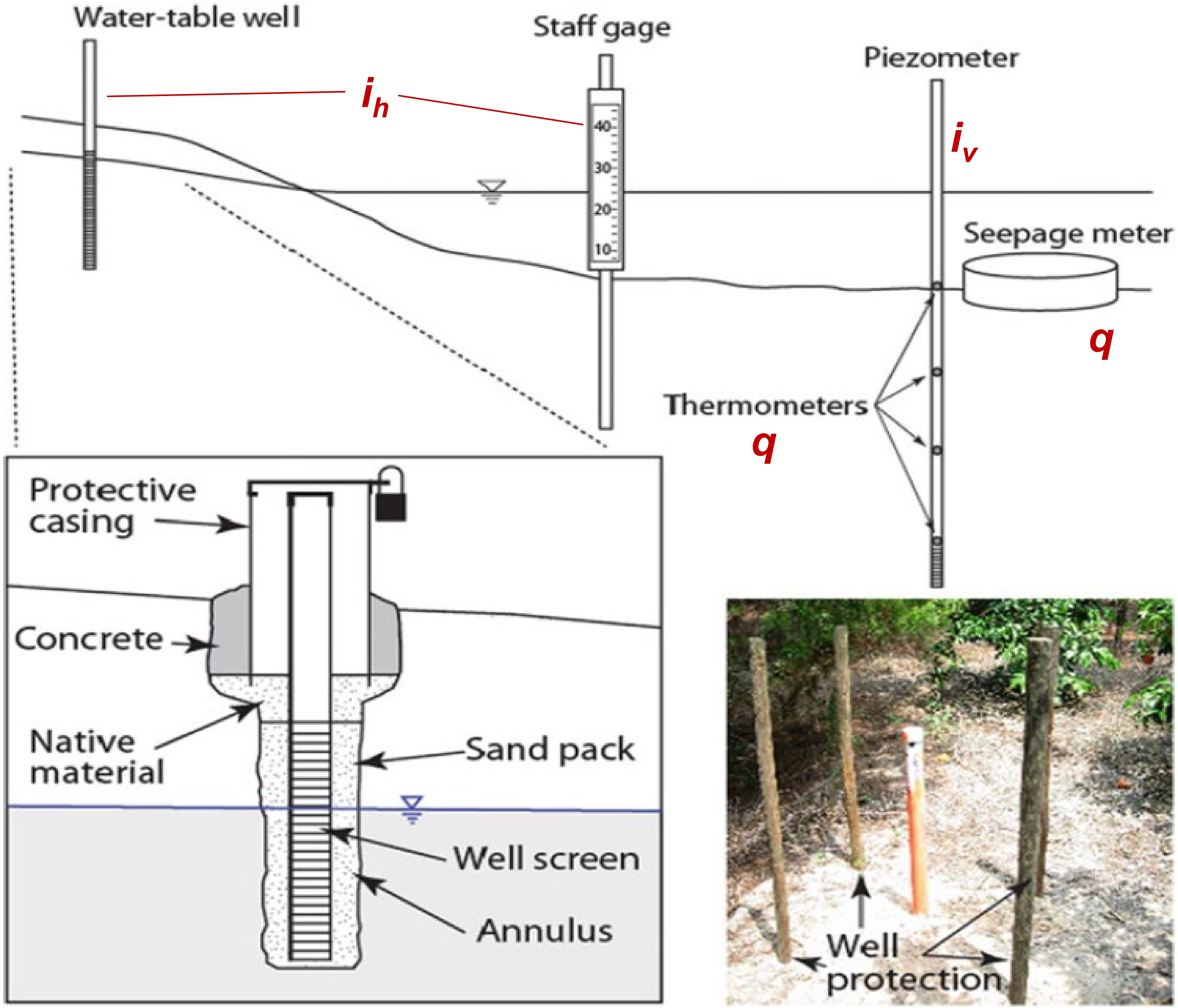
We generally consider a gradient indicating the potential for upward flow as being positive. But that is not always the case. Some hydrogeologists view groundwater flow to surface water as a loss from the groundwater flow domain and, therefore, consider an upward gradient to be negative.




Combining several methods gives us much more confidence in our interpretation. Here, we are measuring vertical hydraulic gradient, horizontal hydraulic gradient, and seepage rate using two different methods.

From Rosenberry and Hayashi, 2013, Chapter 3 – Wetland hydrology

**Fig. 3.41** Typical installation to quantify horizontal and vertical hydraulic gradient, seepage rate, and hydraulic conductivity

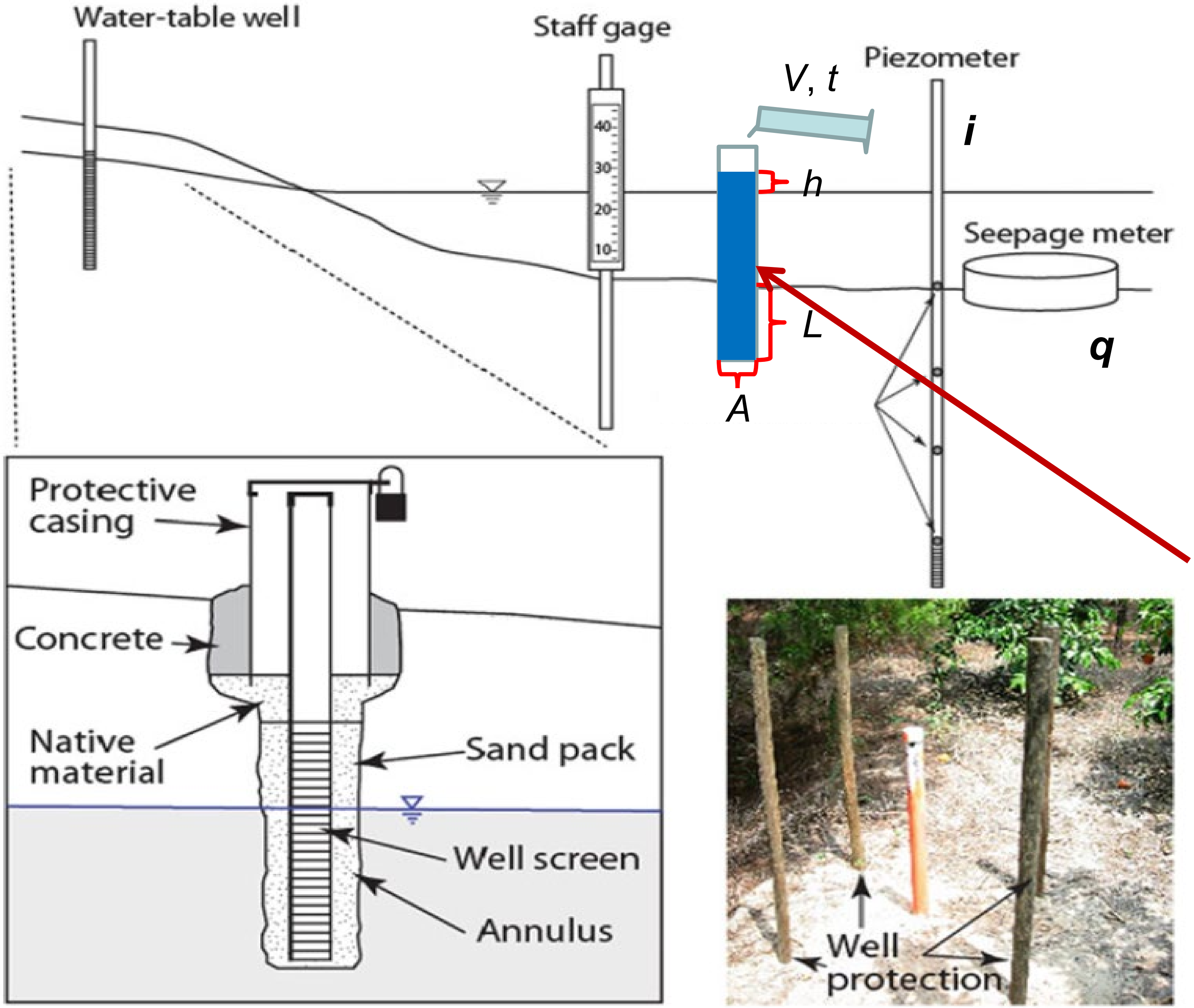



$$K_v = \frac{q}{i}$$

$K_h$ from slug test

If we measure seepage and also vertical hydraulic gradient in essentially the same place, we can calculate $K_v$ integrated over the sediment between the piezometer screen and the sediment-water interface.

**Fig. 3.41** Typical installation to quantify horizontal and vertical hydraulic gradient, seepage rate, and hydraulic conductivity




$$K_v = \frac{q}{i}$$

$K_h$ from slug test

Constant-head permeameter

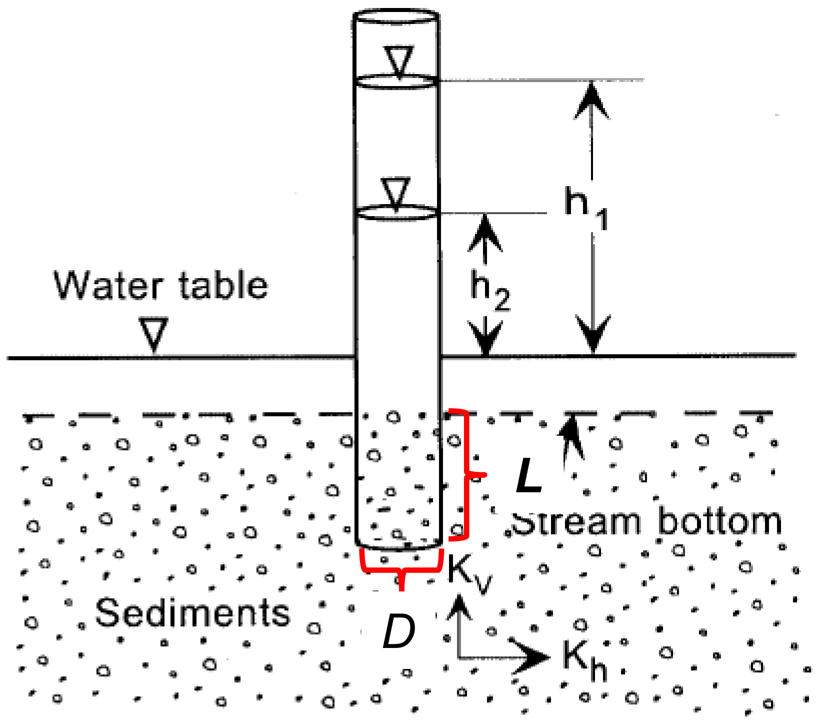
$$K_v = \frac{VL}{Aht}$$

A permeameter gives us another way to measure vertical hydraulic gradient.

**Fig. 3.41** Typical installation to quantify horizontal and vertical hydraulic gradient, seepage rate, and hydraulic conductivity

# Falling-head permeameter

$$K = \frac{L}{t_2 - t_1} \ln \frac{h_1}{h_2}$$



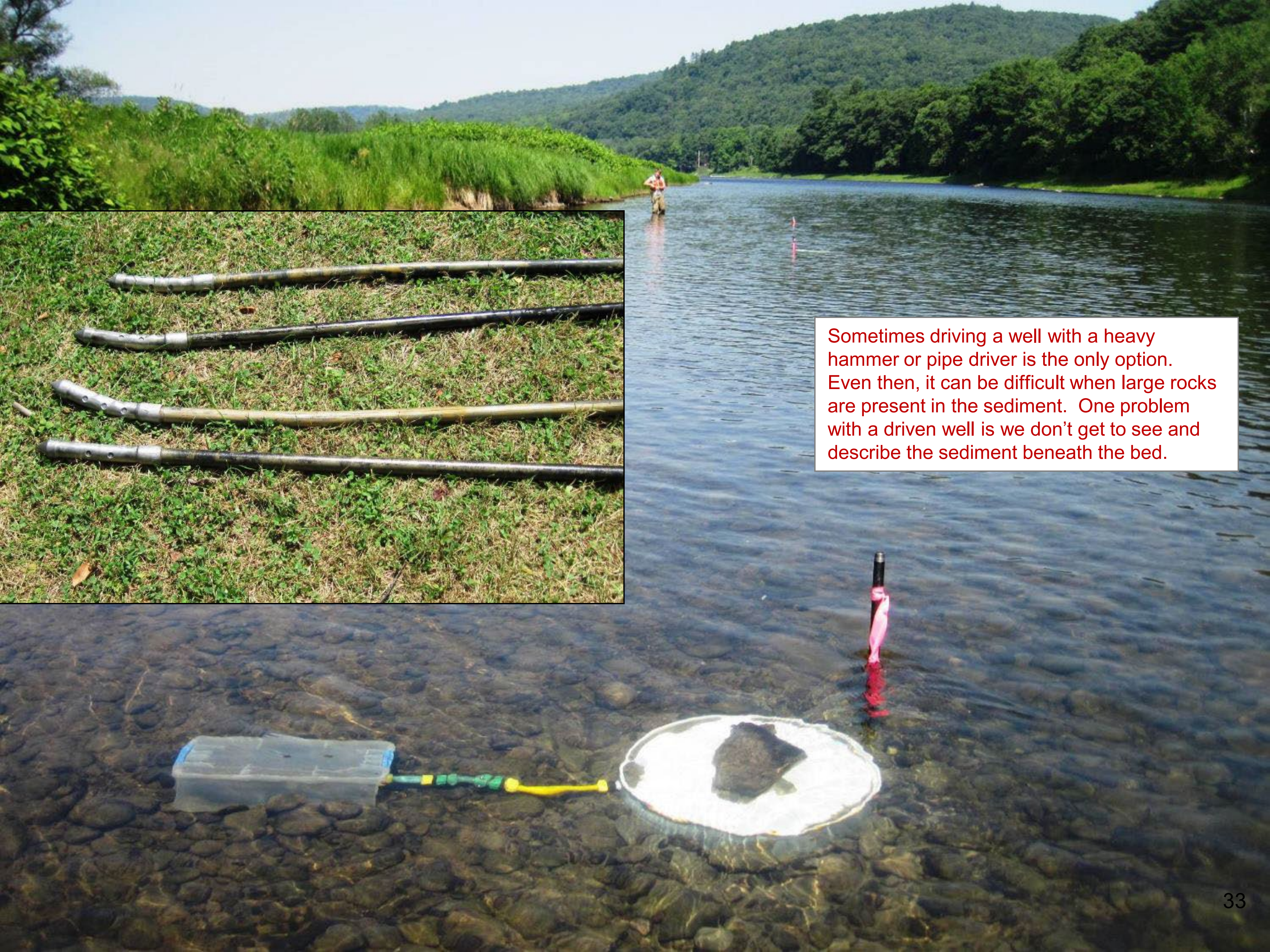

*L* is an approximation. The true solution requires prior knowledge of $K_h$

Good for lower-permeability sediments

Problems

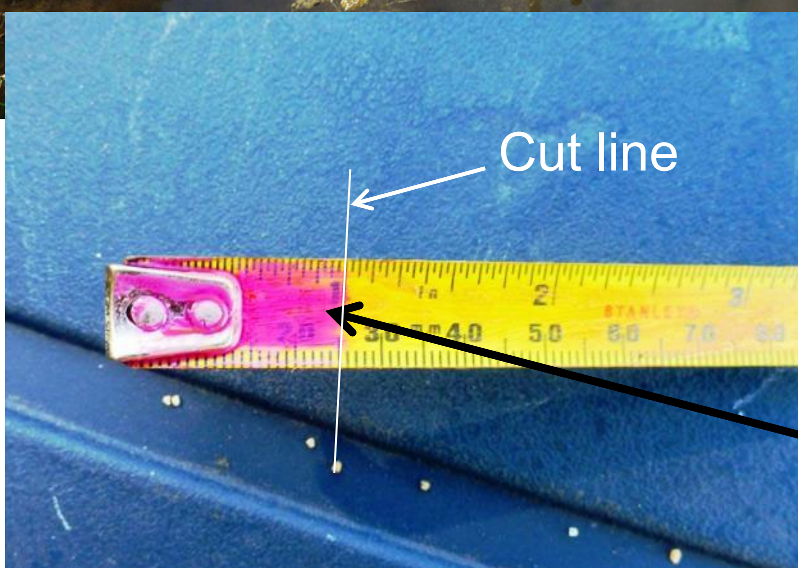
- Underestimates *K* if anisotropy is small (<~20)
- Can compensate with a larger L/D ratio (>~5)
- Difficult to measure for higher-*K* sediments

From Chen, 2000, *Env. Geol*.

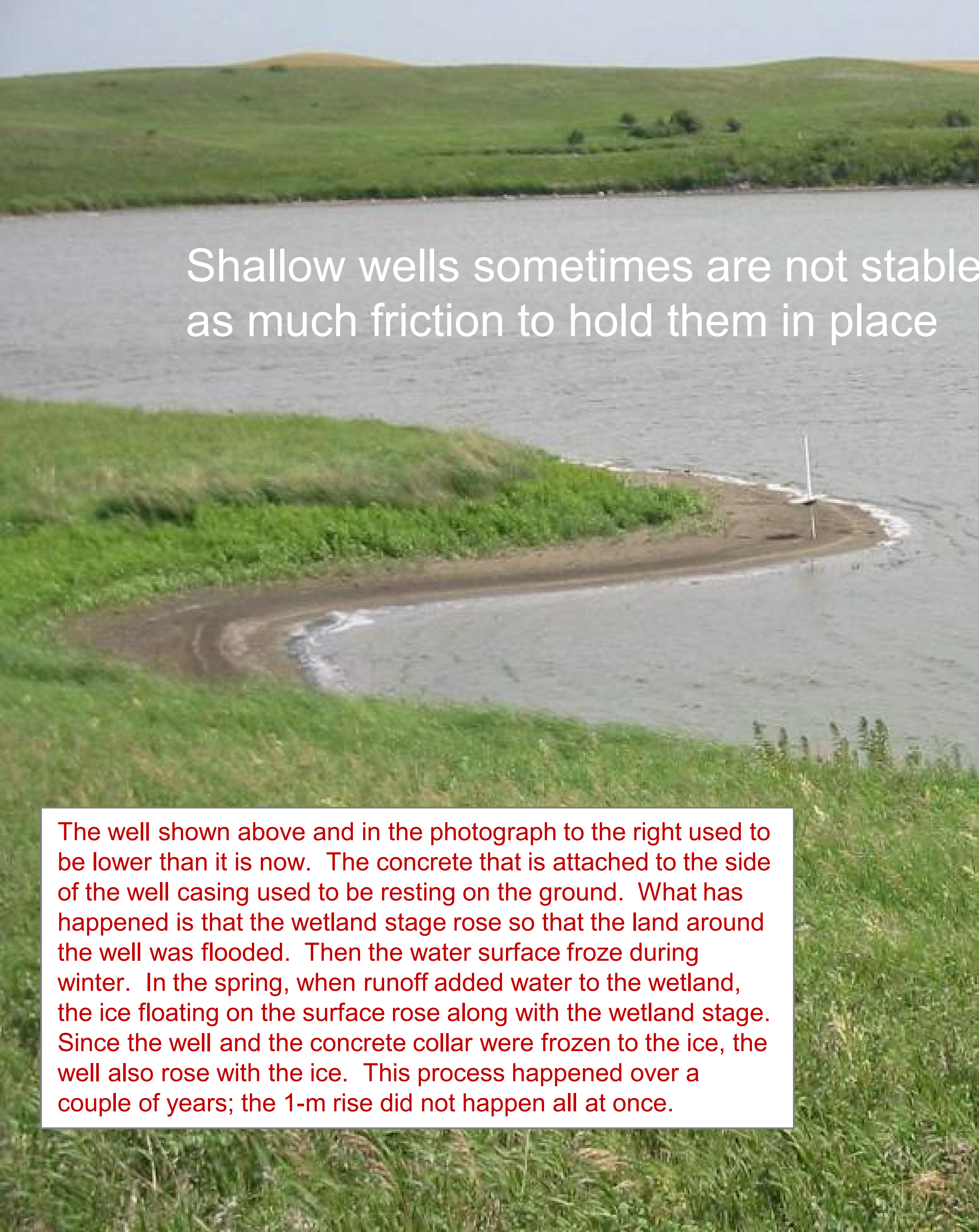


Sometimes driving a well with a heavy hammer or pipe driver is the only option. Even then, it can be difficult when large rocks are present in the sediment. One problem with a driven well is we don't get to see and describe the sediment beneath the bed.





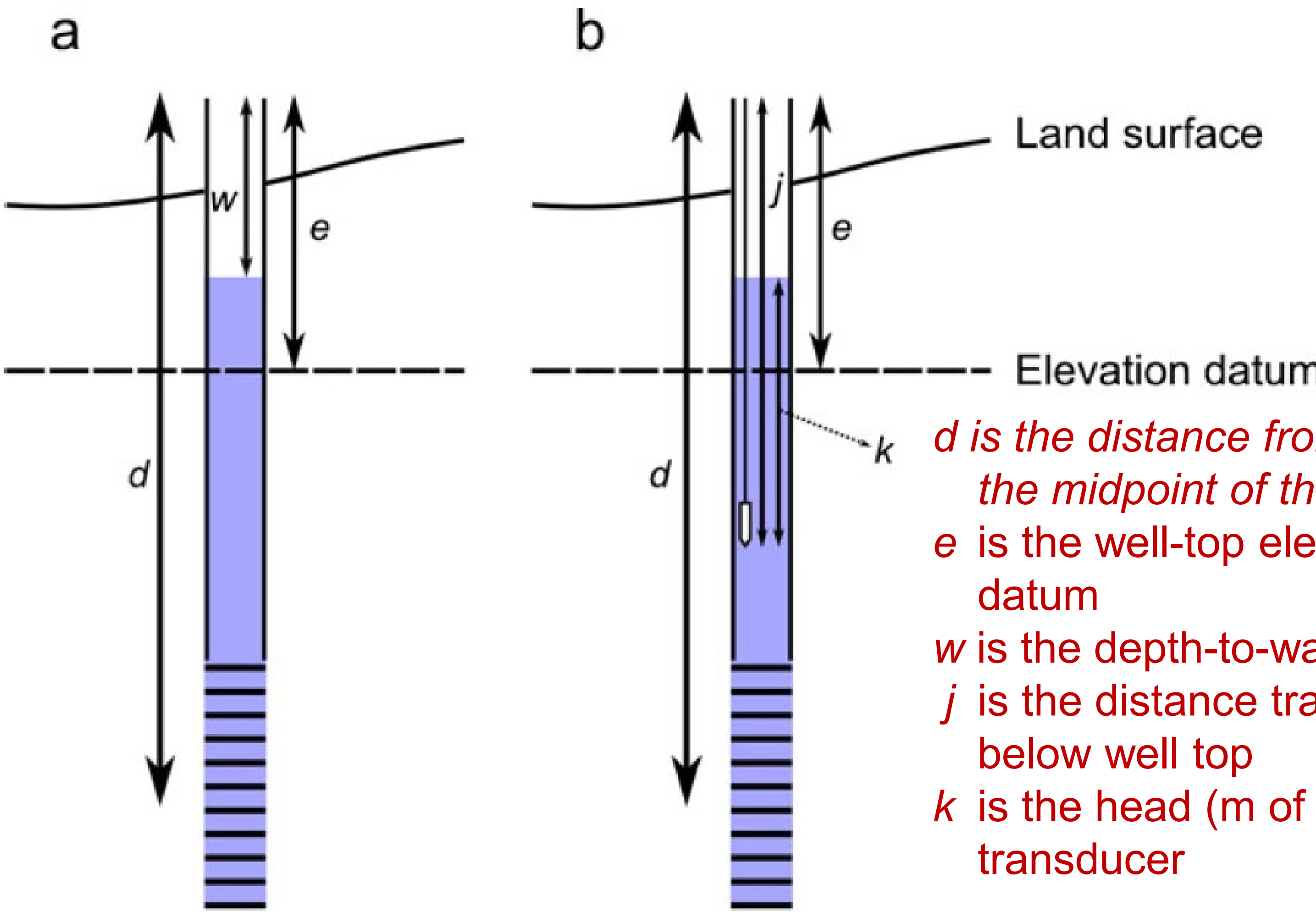

I like Kolor Kut Paste for measuring depths to water



Shallow wells sometimes are not stable – there is not as much friction to hold them in place

The well shown above and in the photograph to the right used to be lower than it is now. The concrete that is attached to the side of the well casing used to be resting on the ground. What has happened is that the wetland stage rose so that the land around the well was flooded. Then the water surface froze during winter. In the spring, when runoff added water to the wetland, the ice floating on the surface rose along with the wetland stage. Since the well and the concrete collar were frozen to the ice, the well also rose with the ice. This process happened over a couple of years; the 1-m rise did not happen all at once.

# Measuring water levels in wells is not trivial




$h = e - w$

$h = e - j + k$

*d* is the distance from the well top to the midpoint of the well screen

*e* is the well-top elevation relative to datum

*w* is the depth-to-water measurement

*j* is the distance transducer is hung below well top

*k* is the head (m of water) above the transducer

But how to get a good value for *j* ? . . .

. . . From $j = w + k$

Post & von Asmuth, 2013, *HJ*

Hydrol. Earth Syst. Sci., 23, 3603–3629, 2019
https://doi.org/10.5194/hess-23-3603-2019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"Further, it takes extraordinary effort to measure hydraulic head gradients <$10^{-3}$ over distances <10 m."

# Error in hydraulic head and gradient time-series measurements: a quantitative appraisal

**Gabriel C. Rau[1,2], Vincent E. A. Post[3], Margaret Shanafield[4], Torsten Krekeler[3], Eddie W. Banks[4], and Philipp Blum[1]**

[1]Karlsruhe Institute of Technology (KIT), Institute of Applied Geosciences (AGW), Karlsruhe, Germany
[2]Connected Waters Initiative Research Centre (CWI), the University of New South Wales (UNSW), Sydney, Australia
[3]Federal Institute for Geosciences and Natural Resources (BGR), Hanover, Germany
[4]National Centre for Groundwater Research and Training (NCGRT), College of Science and Engineering, Flinders University, Adelaide, Australia

**Abstract.** Hydraulic head and gradient measurements underpin practically all investigations in hydrogeology. There is sufficient information in the literature to suggest that head measurement errors can impede the reliable detection of flow directions and significantly increase the uncertainty of groundwater flow rate calculations. Yet educational textbooks contain limited content regarding measurement techniques, and studies rarely report on measurement errors. The objective of our study is to review currently accepted standard operating procedures in hydrological research and to determine the smallest head gradients that can be resolved. To this aim, we first systematically investigate the systematic and random measurement errors involved in collecting time-series information on hydraulic head at a given location: (1) geospatial position, (2) point of head, (3) depth to water, and (4) water level time series. Then, by propagating the random errors, we find that with current standard practice, horizontal head gradients $< 10^{-4}$ are resolvable at distances $\gtrsim 170$ m. Further, it takes extraordinary effort to measure hydraulic head gradients $< 10^{-3}$ over distances $< 10$ m. In reality, accuracy will be worse than our theoretical estimates because of the many possible systematic errors. Re-

This paper, published recently in HESS, presents a great review of the effects of measurement error on hydrogeological interpretation. Note the sentence above in quotes. We often attempt to measure gradients smaller than this without thinking nearly enough about the cumulative measurement errors.

Rau et al., HESS, 2019

# Hydraulic potentiomanometer
(portable well, mini piezometer)

A hydraulic potentiomanometer is basically a portable well that can be used to measure difference in hydraulic head between the wetland and the piezometer screen. We use a manometer to make the head difference easier to measure. You could also make this measurement with just a pipe and well screen driven into the bed. Simply measure the distance from the top of the well pipe to the water level inside the well, and the distance from the top of the well pipe to the surface water, as shown 4 slides earlier.

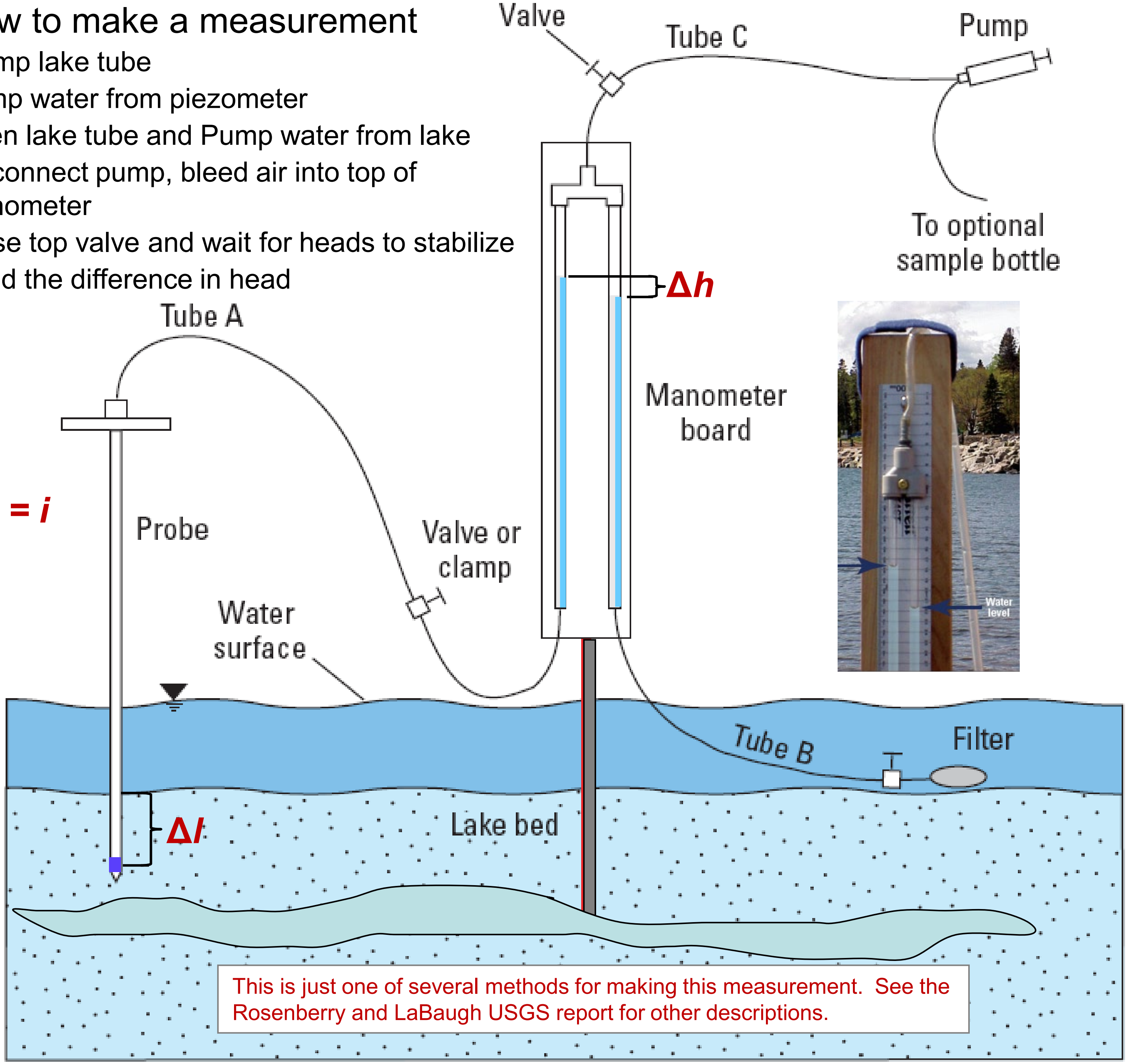

Still, we must try. Here are two designs of a hydraulic potentiomanometer (“mini piezometer”) used to measure vertical hydraulic gradients at a shoreline.

Here are a few types of hydraulic potentiomanometers measuring head gradients at the shoreline of a lake.

# How to make a measurement

1. Clamp lake tube
2. Pump water from piezometer
3. Open lake tube and Pump water from lake
4. Disconnect pump, bleed air into top of manometer
5. Close top valve and wait for heads to stabilize
6. Read the difference in head

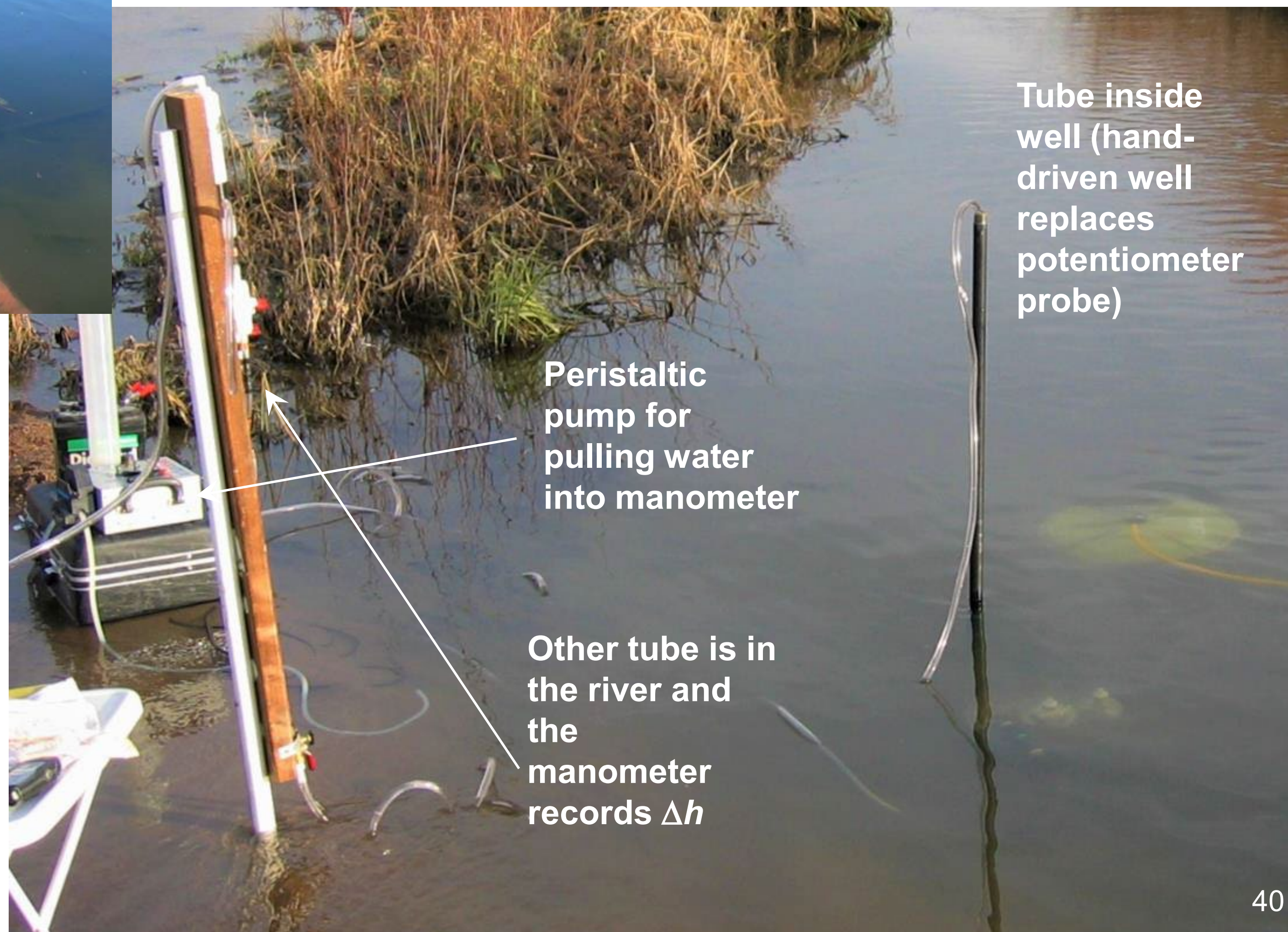
$$\Delta h / \Delta l = i$$




This is just one of several methods for making this measurement. See the Rosenberry and LaBaugh USGS report for other descriptions.

# Other easy ways to measure $\Delta h$ on a vertical axis

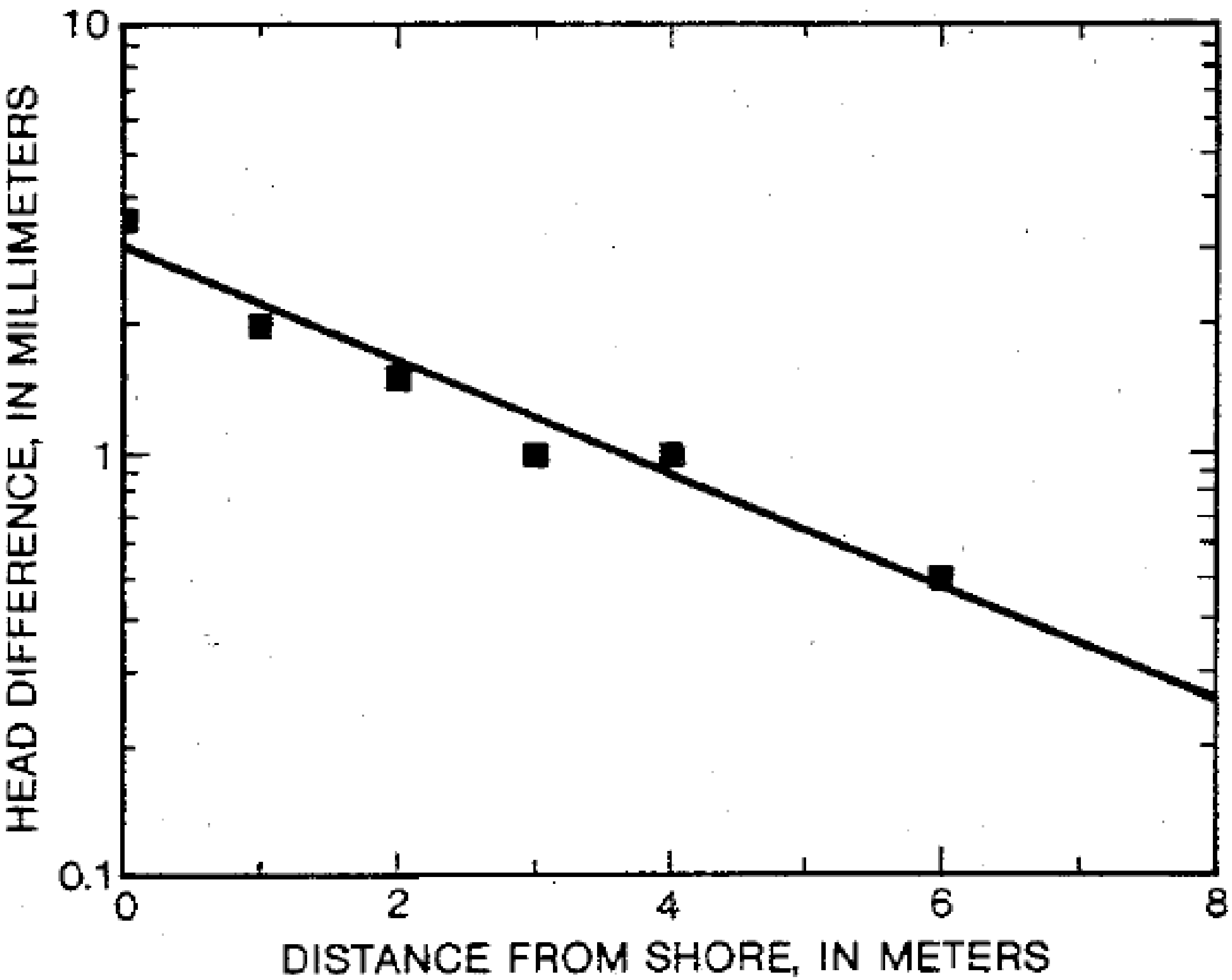

I rarely use the actual hydraulic potentiomanometer setup these days. Instead, I install a shallow well by hand, develop it so it is in good connection with the sediment, and then measure the well level relative to the surface-water level to get $\Delta h$.



# Make measurements with distance from shore

But be sure to insert probe to a consistent depth at all measurement locations

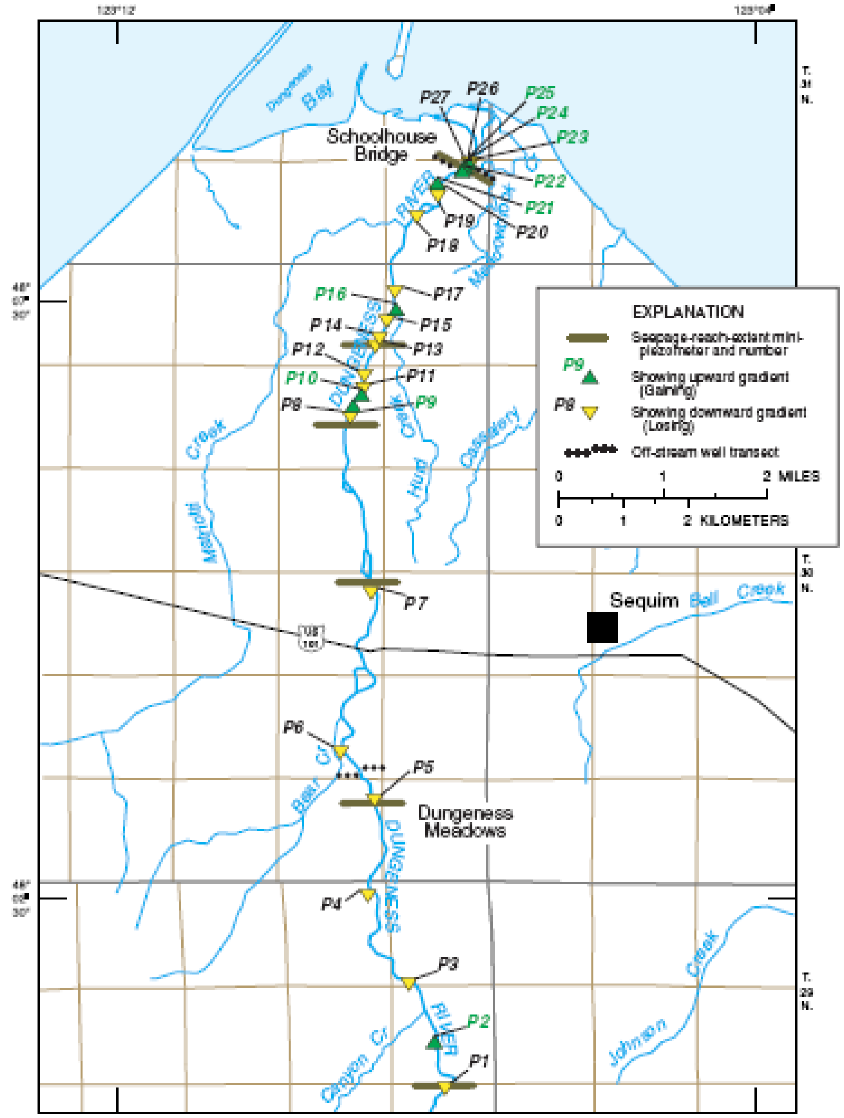
Here we see an exponential decrease in head difference with distance from shore. Since this is what we would expect for a distribution of seepage rate, this indicates that *K* likely is fairly uniform in this area. Data like this are rare, though, indicating sediments are commonly more heterogeneous.




Rosenberry, 1990, NALMS proceedings

# Make measurements with distance downstream

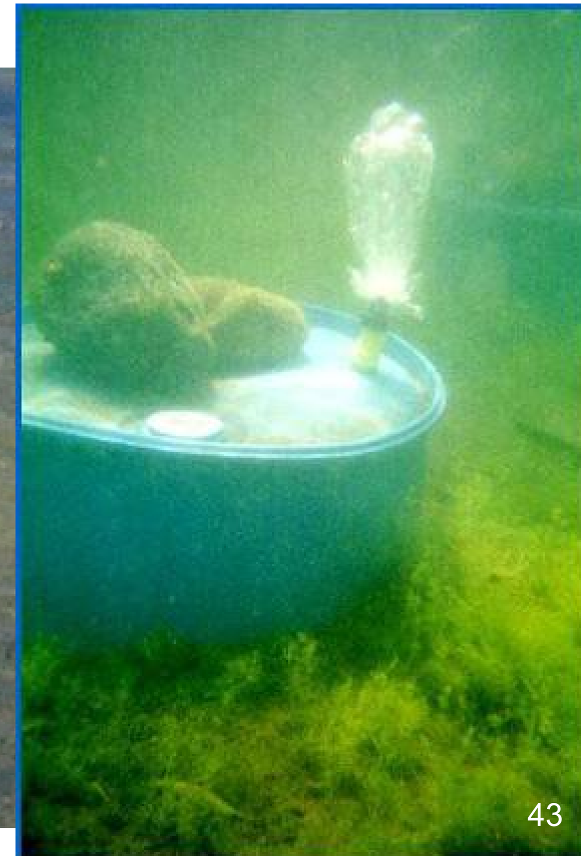
Seepage run gives $\Delta Q$, potentiomanometer gives gradient, can calculate streambed $K$


Simonds & Sinclair, WRIR02-4161

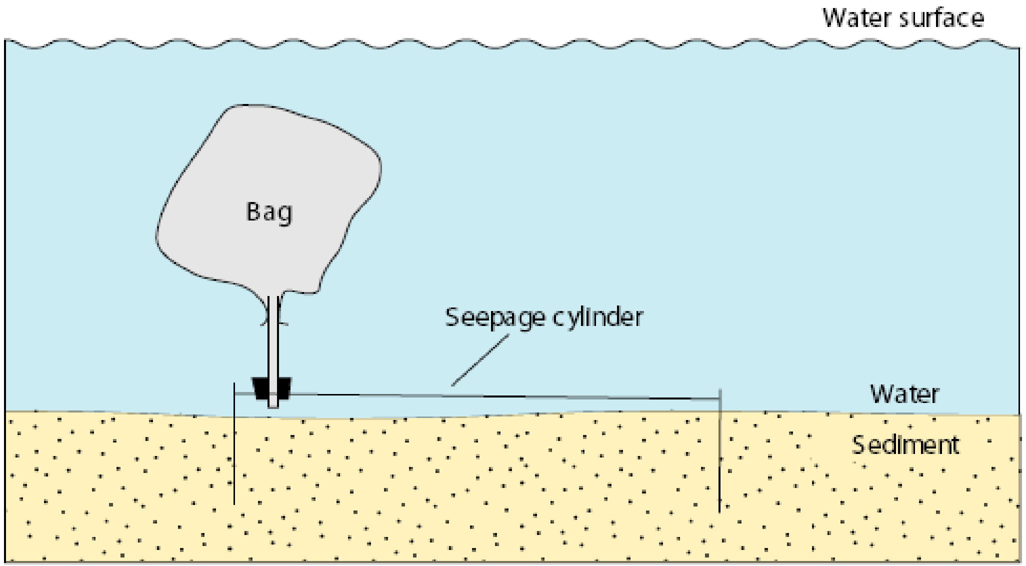
# Seepage meters

- Direct measurement of flux
- Measure flows from ~0.1 to ~500 cm/d (10$^{-8}$ to 5x10$^{-5}$ m/s)
- Modified versions can measure down to ~0.00001 cm/d or up to 5000 cm/d or more



And now let's talk about my favorite instrument for measuring flow between groundwater and surface water. I like this device for several reasons, first that it makes a direct measurement of Q, the only device that I know of that can do that.

Another great thing about this method is it is very simple and very inexpensive.




# Lee-type seepage meter

Lee, D.R., 1977, A device for measuring seepage flux in lakes and estuaries: *Limnology and Oceanography*, v. 22, no. 1, p. 140-147.

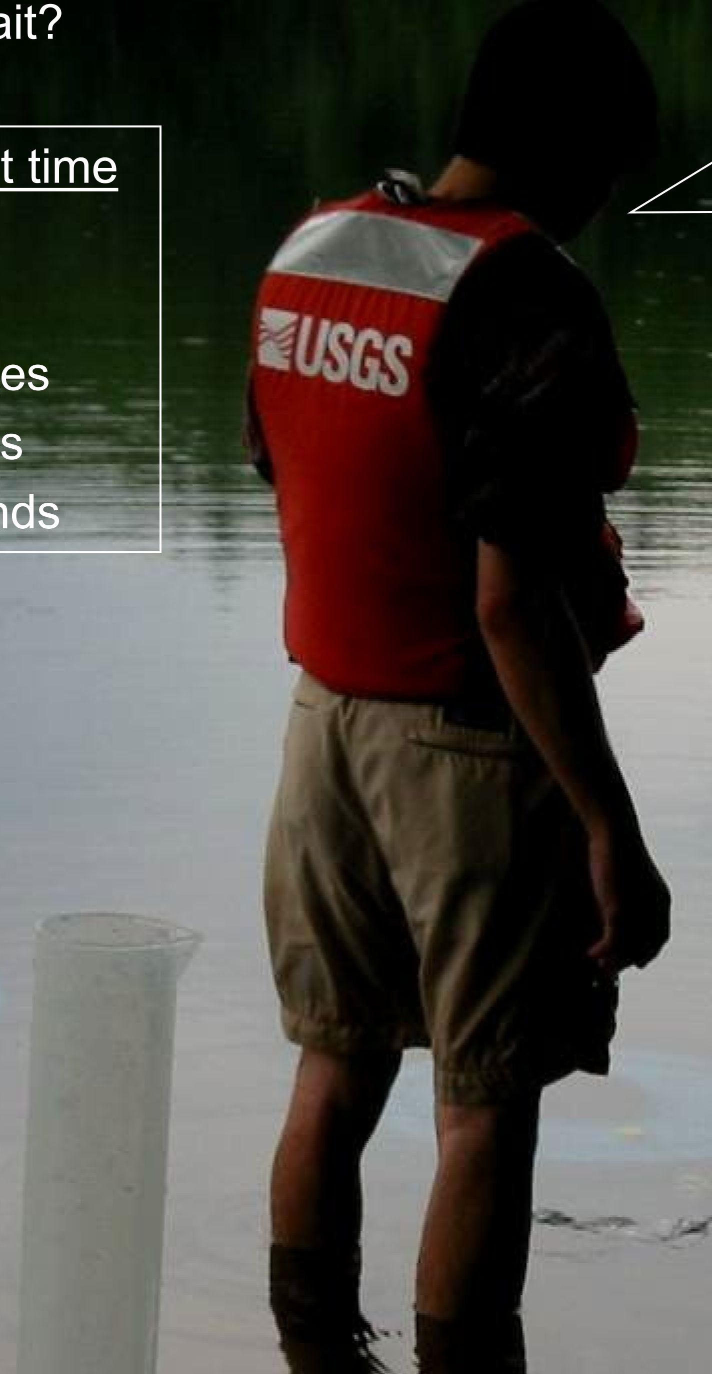
## How to make a measurement

1. Place 500 to 1000 ml of water inside bag
2. Attach bag to cylinder and record time
3. Wait
4. Remove bag from cylinder and record time
5. Determine $\Delta V/\Delta t$ (ml/min) or $cm^3$/sec
6. Divide by area covered by meter to get cm/d or m/s (flux per unit area)

## How long do we have to wait?

| Flux rate (cm/d) | Bag-attachment time |
|---|---|
| 0.1 | 0.25 to 2 days |
| 1 | 1 to 10 hours |
| 10 | 10 to 60 minutes |
| 100 | 1 to 10 minutes |
| 1000 | 30 to 90 seconds |

When will it ever end!?



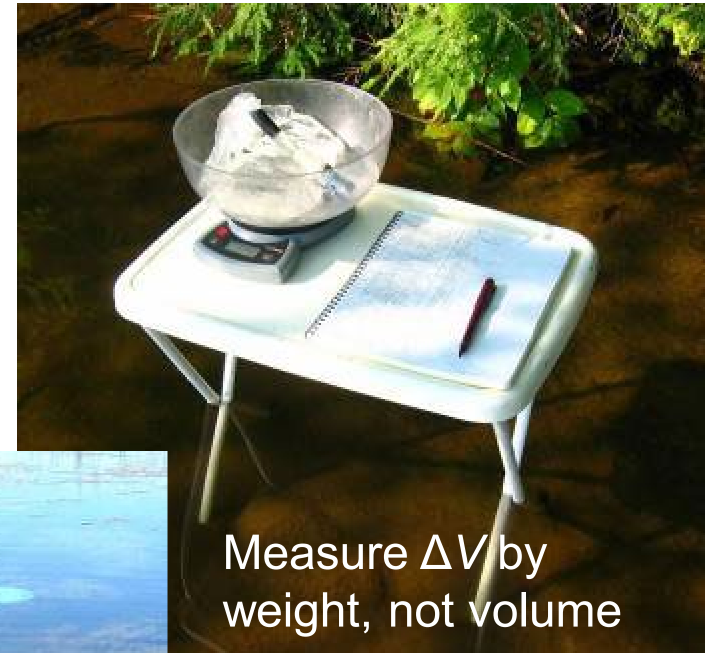
- Larger errors for flows slower than 0.1 cm/d
  - Bag resistance becomes significant
- Friction head loss becomes significant for flows greater than ~200 to 400 cm/d
  - Need to use larger-diameter plumbing and tubing

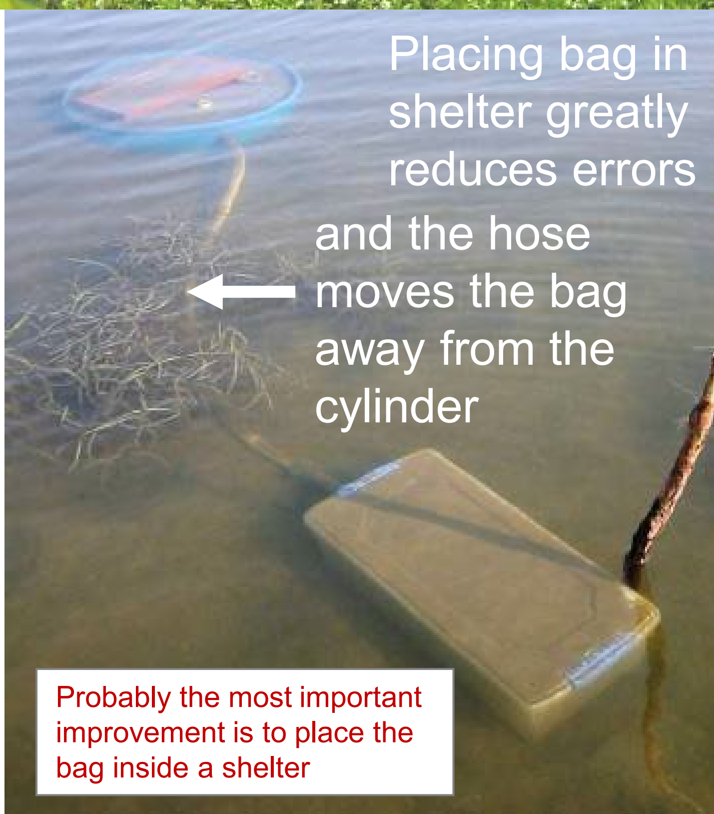
Bag-attachment time depends on the seepage rate. If the time is too short, we have a volume change that is smaller than the measurement error. If the time is too long, the bag will be full or empty. And we don’t know the seepage rate when we first visit a site. The first measurement or two usually is a best guess regarding how long to leave the bag attached. I like to start with 20 to 30 minutes. If the bag is full or empty when removed, then the next measurement duration should be shorter. If there is very little change in volume, then the next measurement duration should be longer.

# Some improvements over the decades




# More improvements



# Sources of error

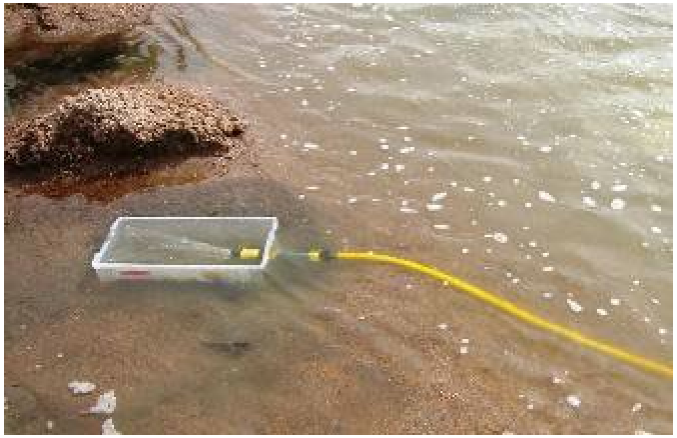
- Incomplete seal, unstable cylinder (a common and substantial error)
- Insufficient time between cylinder installation and first measurement
- Leaks
- Improper bag attachment, bag resistance, moving water
- Measurement error (reading/writing the wrong number)
- Flexible seepage cylinder
  - Cylinder can deform during installation and later “pop” back into shape
- Insufficient or excessive bag-attachment time
  - Bag either sucked dry or plump
  - Iterative solution (times can vary from ~30 seconds to days)
- Accumulation of trapped gas
  - Problem in organic sediments, especially during long measurements
  - Frequent venting of cylinder is the solution
- Incorrect seepage-meter coefficient
- Poor characterization of spatial heterogeneity
- Wind and waves (solution is to place bag inside a shelter)
- Surface water flow too fast or too slow for the meter design

Early users of these devices were not aware of many of these sources of error and some of the published data likely were in error. Therefore, these devices got a bad reputation among some scientists. It has taken decades to identify these sources of error, determine ways to eliminate them, and restore the reputation of these devices in the scientific community.

# Seepage bags


Void-fill bags, Inflatable Packaging, Inc.

Many of the errors were associated with the bag. Best results are produced by placing the bag in a bag shelter, using large-diameter tubing, and using a thin-walled, flexible bag.




Here are a couple of bag shelters with bags attached inside and in the process of making a seepage measurement.

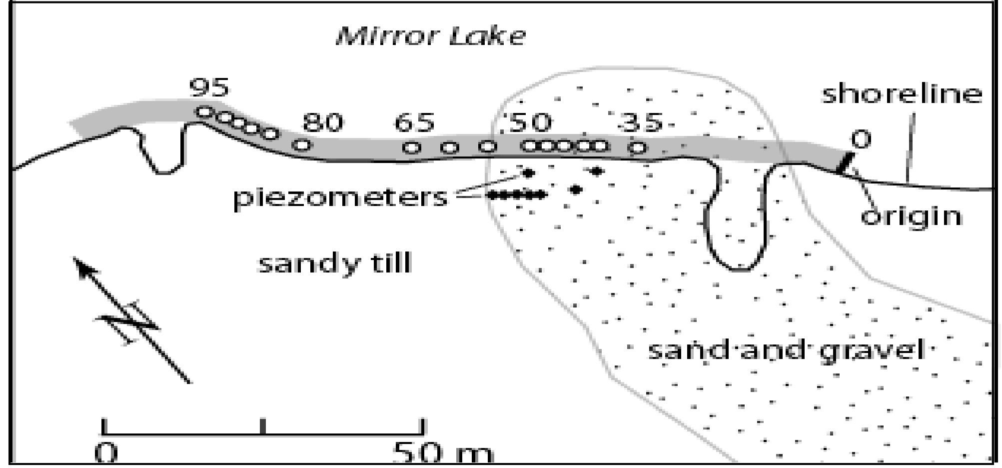
# It takes a lot of walking around to measure seepage

Speaking of walking, just don't do it!
Well, OK, maybe you do have to some walking to measure seepage. But don't forget the take-home message that you will get from the next slide.

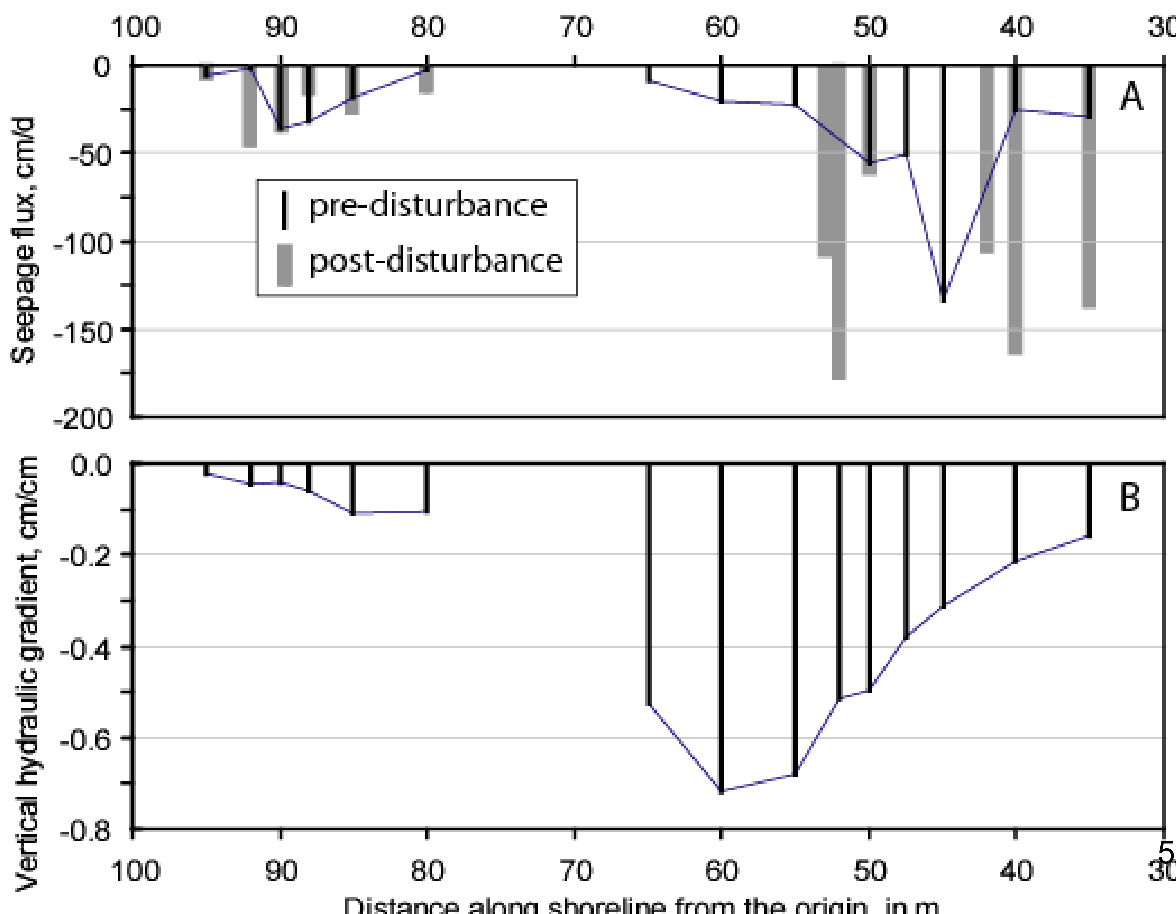

It looks like we created a highway

Rosenberry et al., 2010, *WRR*

Seepage increased by factors of 2.6 to 7.7 based on comparing pre- to post-disturbance measurements

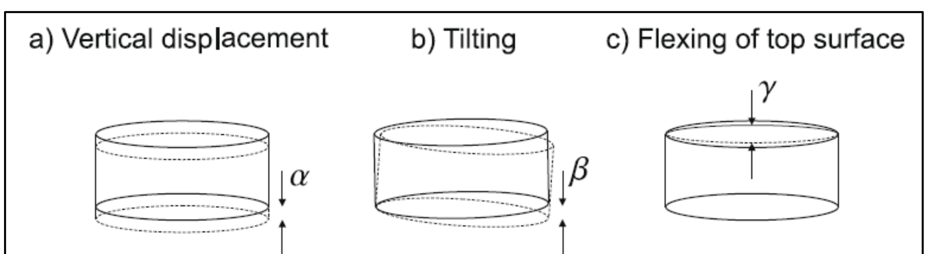



A basic problem with scientific observation in general is that we often affect the process we are trying to quantify with our measurement of that process. That certainly is the case with manual measurements of seepage. This paper indicated that we disturbed a thin layer of silt deposited on the bed that was only a few millimeters thick. This disturbance caused an increase in seepage by a factor of about 2 to nearly an order of magnitude. One wonders how many other data published in the literature, including several papers published by yours truly, are greatly in error because we were not aware of this disturbance.




Rosenberry et al., 2010, *WRR*

•Wind and waves


Smith et al., 2009, *AWR*


Better to quit for the day than collect bad data.

Even with the bag placed in a shelter, waves can still corrupt a seepage meter measurement. One rogue wave could in a few seconds cause huge hydraulic gradients that could greatly change the volume of water in the seepage bag. The solution is to make longer measurements and hope that any short-term large (or small) fluxes are time-averaged.

# Everything in the previous 8 slides is presented here, and much more. The latest, greatest seepage-meter summary

We used to have quite a few more slides on seepage meters for this course, but this two-part course, but this two-part summary covers that information and more.

Earth-Science Reviews 204 (2020) 103167

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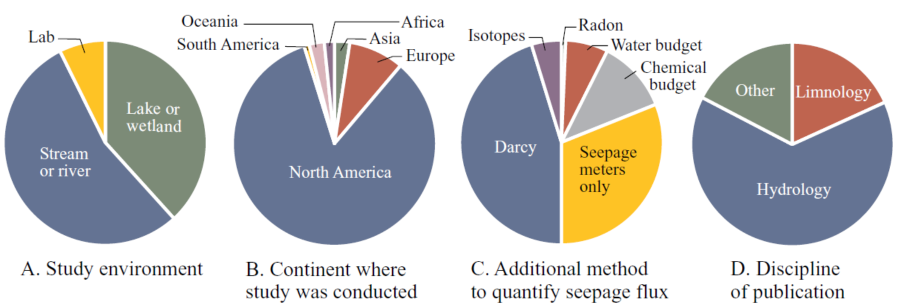
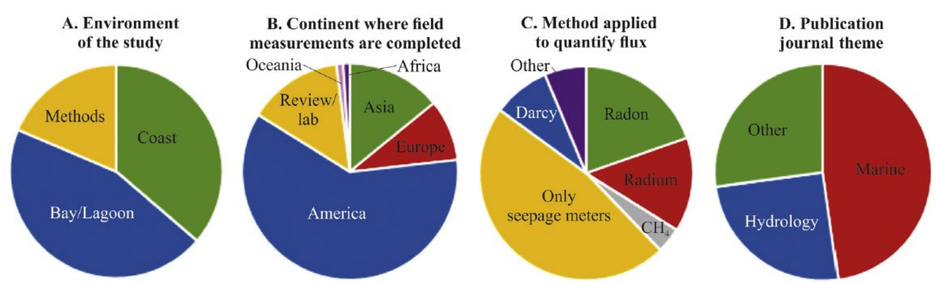
Carlos Duque[a], Christopher J. Russoniello[b], Donald O. Rosenberry[c,*]
[a] WATEC, Department of Geoscience, Aarhus University, Aarhus, Denmark
[b] Department of Geology and Geography, West Virginia University, Morgantown, WV, USA
[c] U.S. Geological Survey, Lakewood, CO, USA

Summarizing the seepage-meter literature to date was an interesting exercise. The emphasis has shifted from lakes to fluvial systems, from primarily freshwater to marine margins, from just using seepage meters to using them in combination with other methods and devices.

And the seepage-meter science continue to evolve. Meters have become more efficient over the years. However, as the next few slides show, efficiency isn't just related to sensor design.




Fresh water




Saline and brackish water

# Incorrect seepage-meter coefficient – a source of bias

**Table 1 Reported correction factors to adjust seepage-meter flow rates to actual rates.**

| Citation | Correction factor |
|---|---|
| (Erickson, 1981) | 1.43 (flow from ground water to surface water) |
| (Erickson, 1981) | 1.74 (flow from surface water to ground water) |
| (Cherkauer and McBride, 1988) | 1.6 |
| (Dorrance, 1989) | 1.61 |
| (Asbury, 1990) | 1.11 |
| (Belanger and Montgomery, 1992) | 1.3 |
| (Murdoch and Kelly, 2003) | 1.25 |
| (Rosenberry, 2005) | 1.05 |

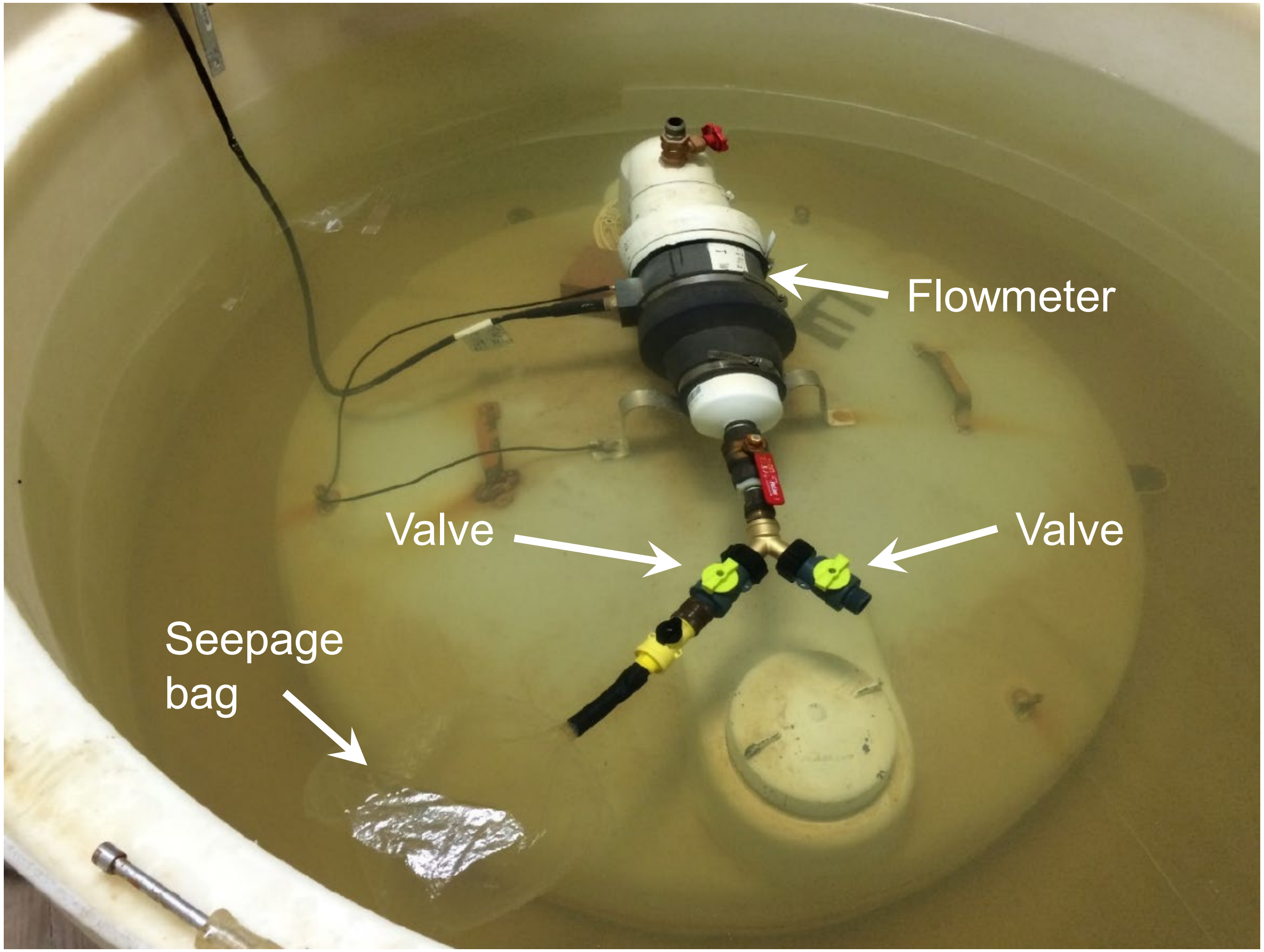
Rosenberry and Menheer, SIR 2006-5053

Solution? -- Use thin bags and large-diameter tubing

Note that seepage meters are becoming more efficient with improvements in design. In general, correction coefficients have become smaller over time. One exception is Asbury, 1990. Clyde Asbury was ahead of his time and likely would have done great things in science if not for his untimely death.

But wait! It turns out seepage meter efficiency is not a constant value in highly permeable settings (wave-washed beaches, regularly mobile fluvial sediments)

Here we relate seepage flowing through an automated flowmeter with no bag attached, to seepage measured by the flowmeter when a bag is attached. If we had 100 percent bag efficiency, there should be no change in flow whether the bag is attached or not. But that never happened. Seepage was always reduced when a bag was attached.

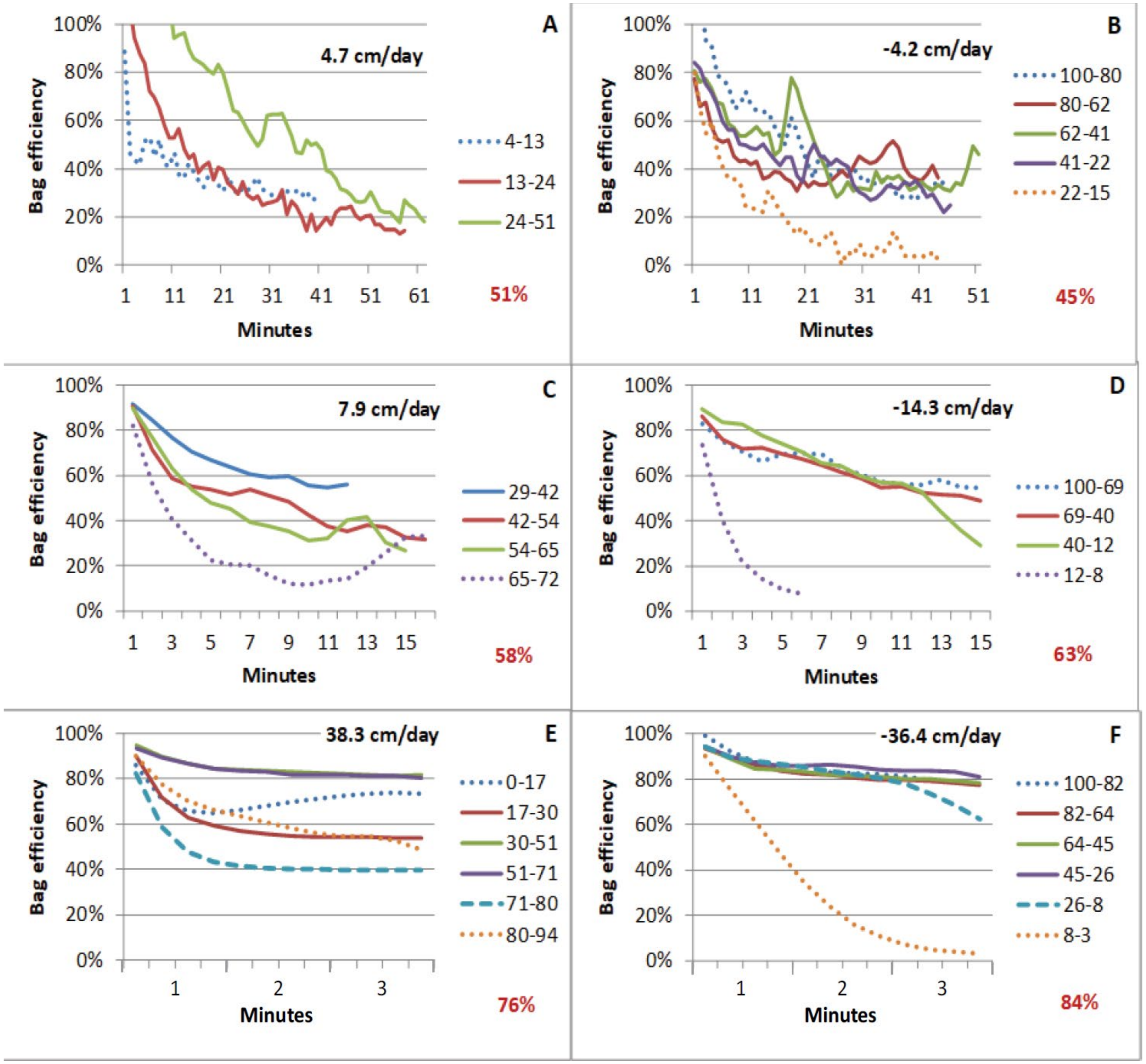



Rosenberry et al., 2020, Water

# Seepage-bag efficiency depends on bag fullness, duration of attachment, direction of seepage, and depth of seepage cylinder insertion

Here we're showing bag efficiency each minute of each bag attachment during upward (A,C, E) and downward (B, D, F) seepage during different ranges of bag fullness, and for slow, medium, and faster seepage rates. The percentages in red are average efficiencies for the bags operated between 1/5 and 2/3 full, indicating that efficiency improves for faster seepage rates.

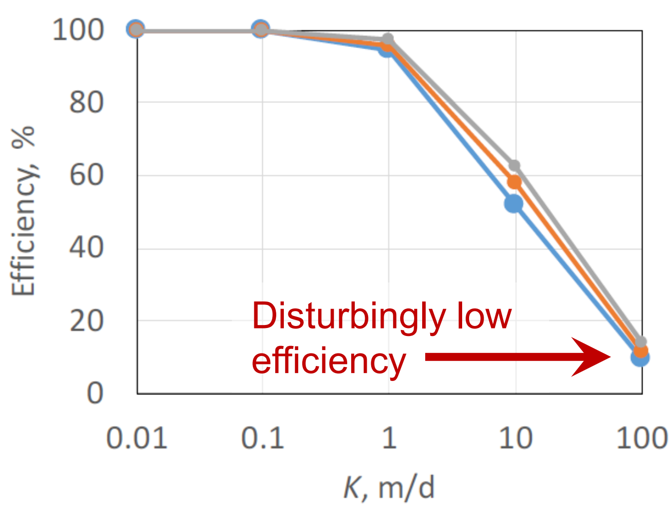
There's a lot of information here. Efficiency varies most for the slowest seepage rates, likely because smaller seepage forces cannot overcome small changes in bag resistance to inflation or deflation. Some efficiencies are well over 100 percent shortly after the bag valve was opened to start a measurement, likely because the bag was under small tension or compression.

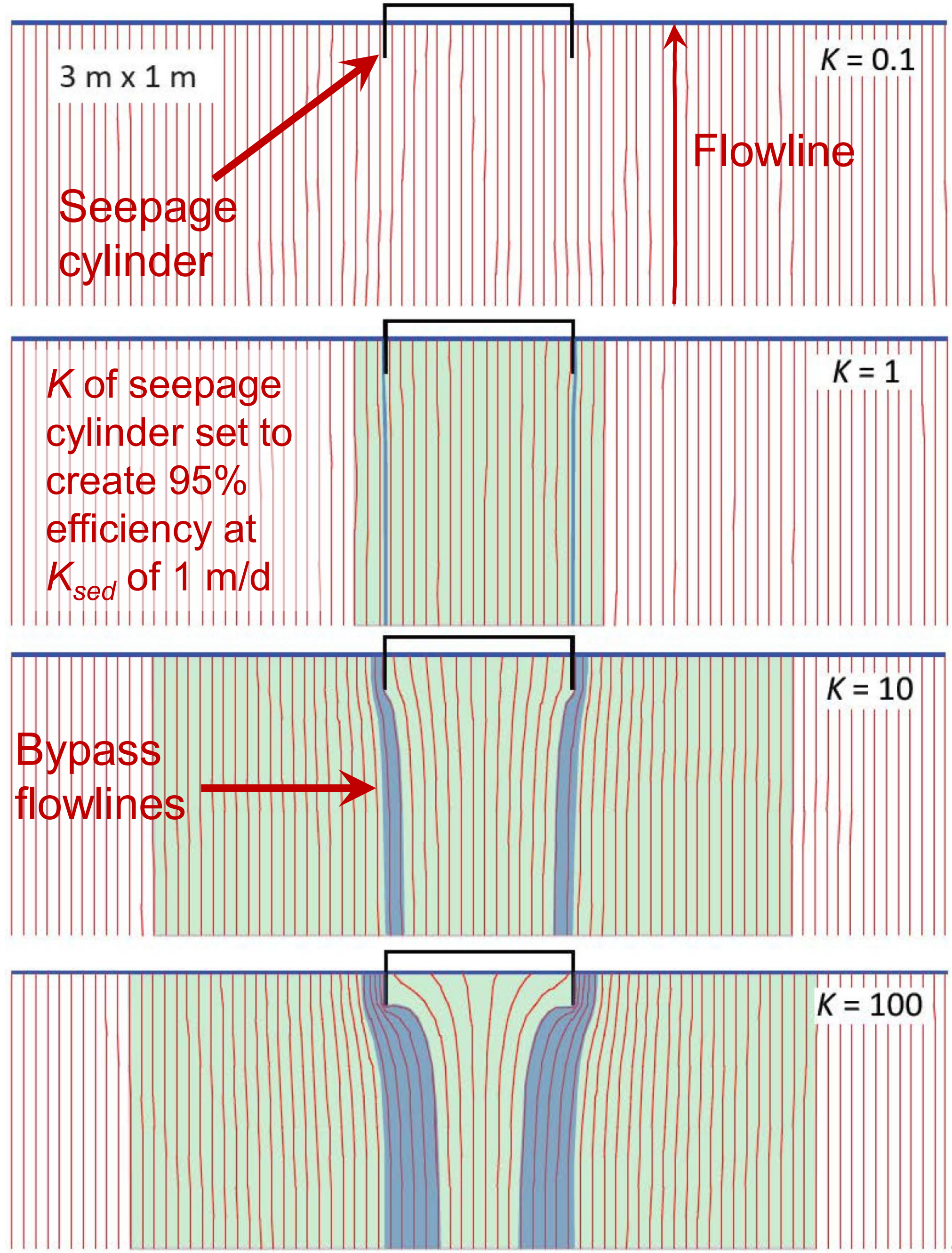



Rosenberry et al., 2020, Water

# Simulated flowlines show bypass flow that reduces efficiency with increasing *K*

Simulated flow is from high constant head at the bottom of the domain to lower constant head representing the sediment-water interface at the top of the domain. Blue shading indicates flowlines that would have ended inside of the seepage cylinder but instead are diverted beyond the edge of the cylinder due to increased resistance of flow inside the cylinder. Green shading indicates the portion of the flow domain where flowlines are visibly diverted due to the presence of the seepage cylinder. Simulated hydraulic conductivity ranges from 0.1 to 100 m/day.




Rosenberry et al., 2020, Water

# Why not just replace the bag with a flowmeter?

Why not do away with the bag completely? The next few slides demonstrate ways of doing just that by replacing the bag with a flowmeter capable of measuring very slow flows typical of seepage.

# Heat-pulse seepage meter 1993

Taniguchi sensei changed the way we view seepage. We used to think that seepage did not vary much in time. Thanks to Taniguchi san's many papers, we now know that is not true.

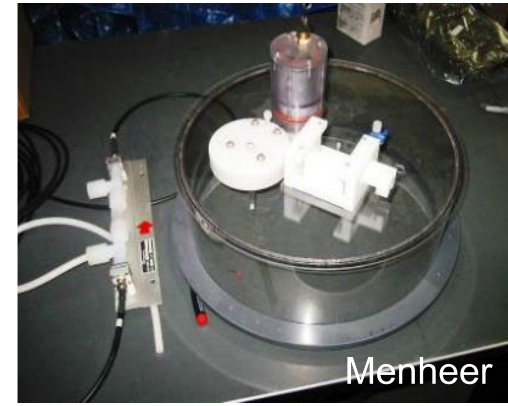
Taniguchi, M., and Y. Fukuo. 1993. Continuous measurements of ground-water seepage using an automatic seepage meter. *Ground Water* 34 (4):675-679.

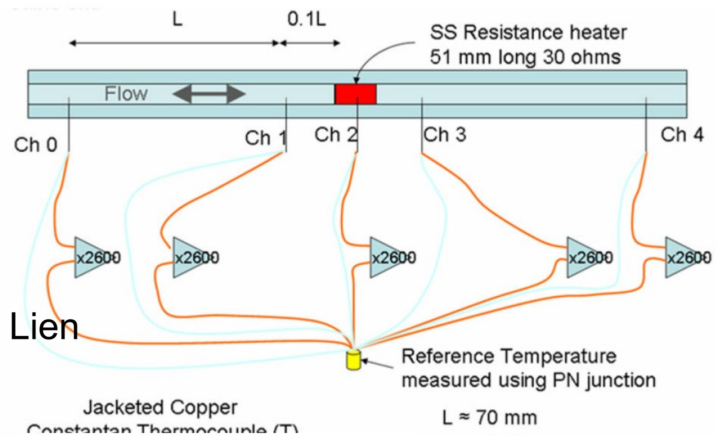
***Newer version discussed in 2001 Journal of Ground Water Hydrology***

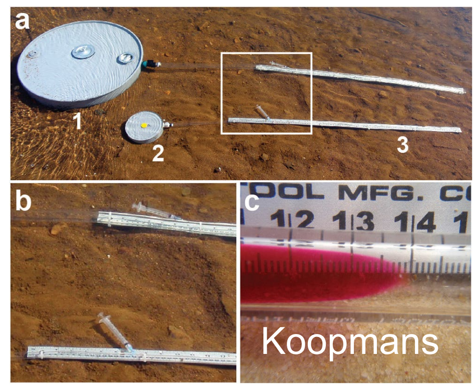
# Other automated seepage meters

Details in Rosenberry et al., 2020, *Earth Sci. Rev.*

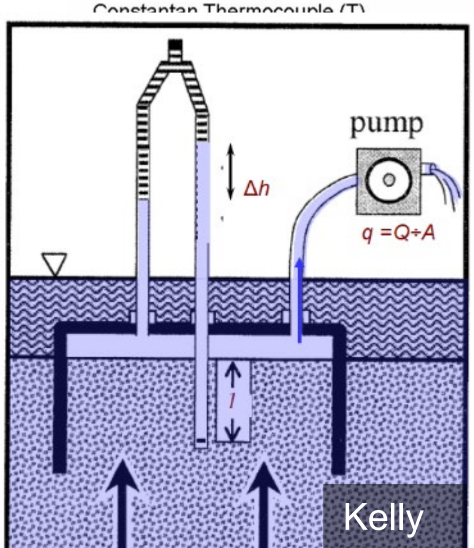
Paulsen, 2001, *Groundwater* – Ultrasonic flowmeter – led to Coastal Monitoring's patented "Ultraseep
Sholkovitz et al., 2003, *L&O-Methods* – Dye dilution
Kelly & Murdoch, 2003, *Groundwater* – "Piezoseep"
Menheer, 2004, USGS SIR – Ultrasonic flowmeter
Rosenberry and Morin, 2004, *Groundwater* – Electromagnetic flowmeter
Lien, 2006, EPA – modified heat-pulse flowmeter
Koopmans & Berg, 2011, *WRR* – Dye displacement
Zhu et al., 2015, *Env. Mon. & Assess* – Revised heat-pulse
**Solomon et al., 2020, *WRR* – "Tube-seep" dh/dt revision of Bouwer method from 1960s**

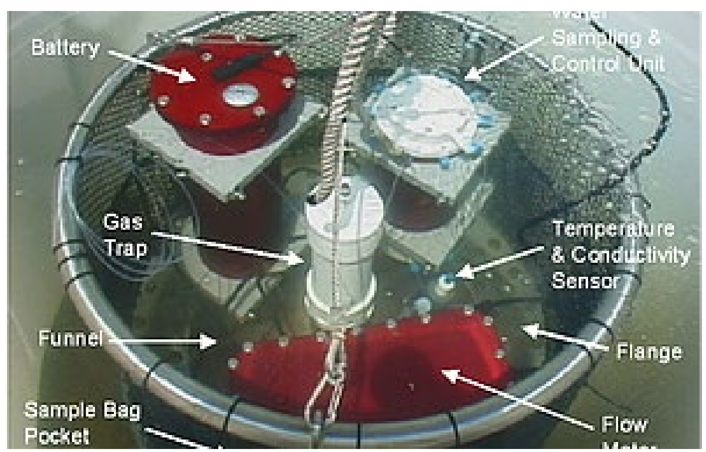


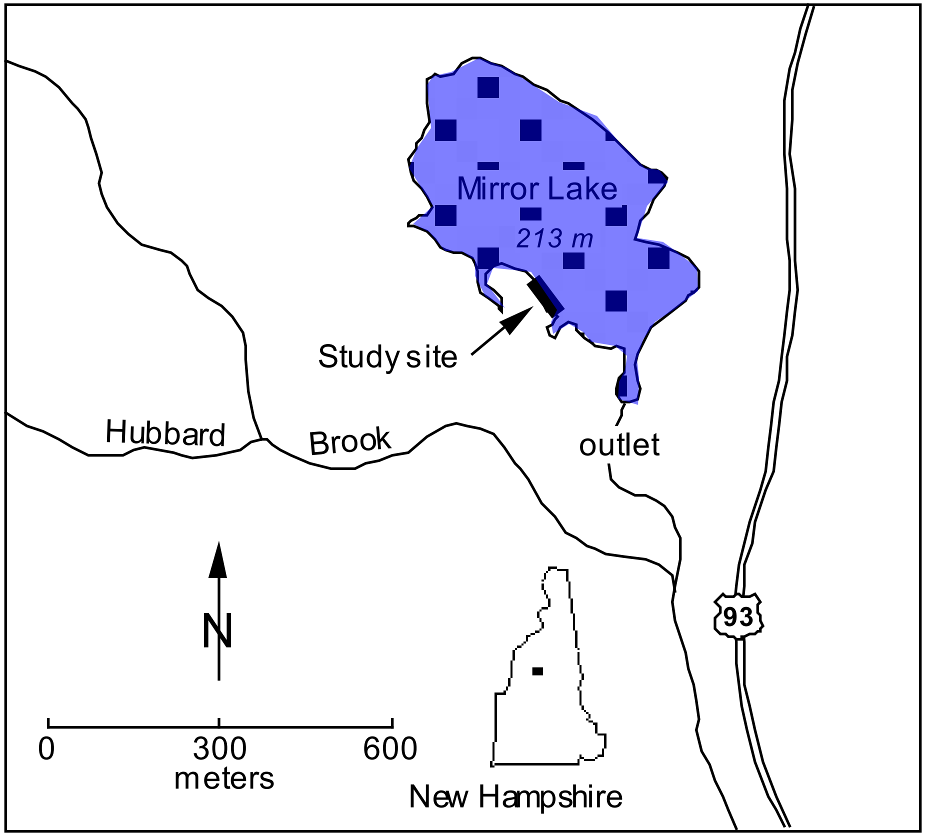



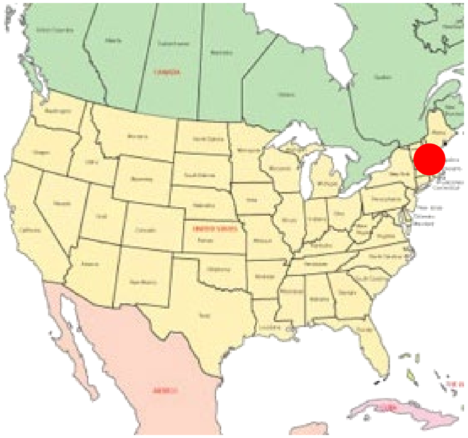





Ultraseep



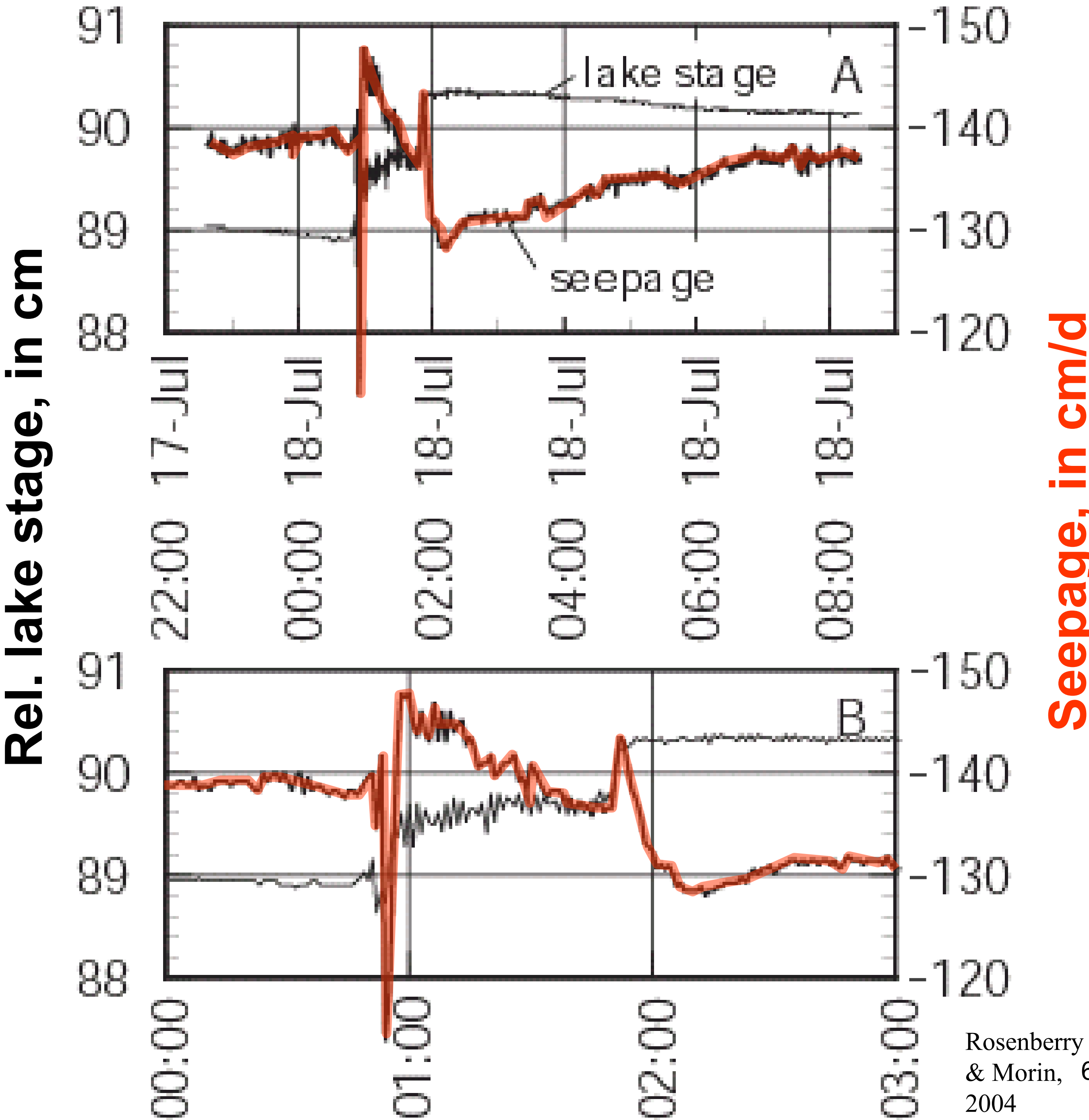



Electromagnetic seepage meter

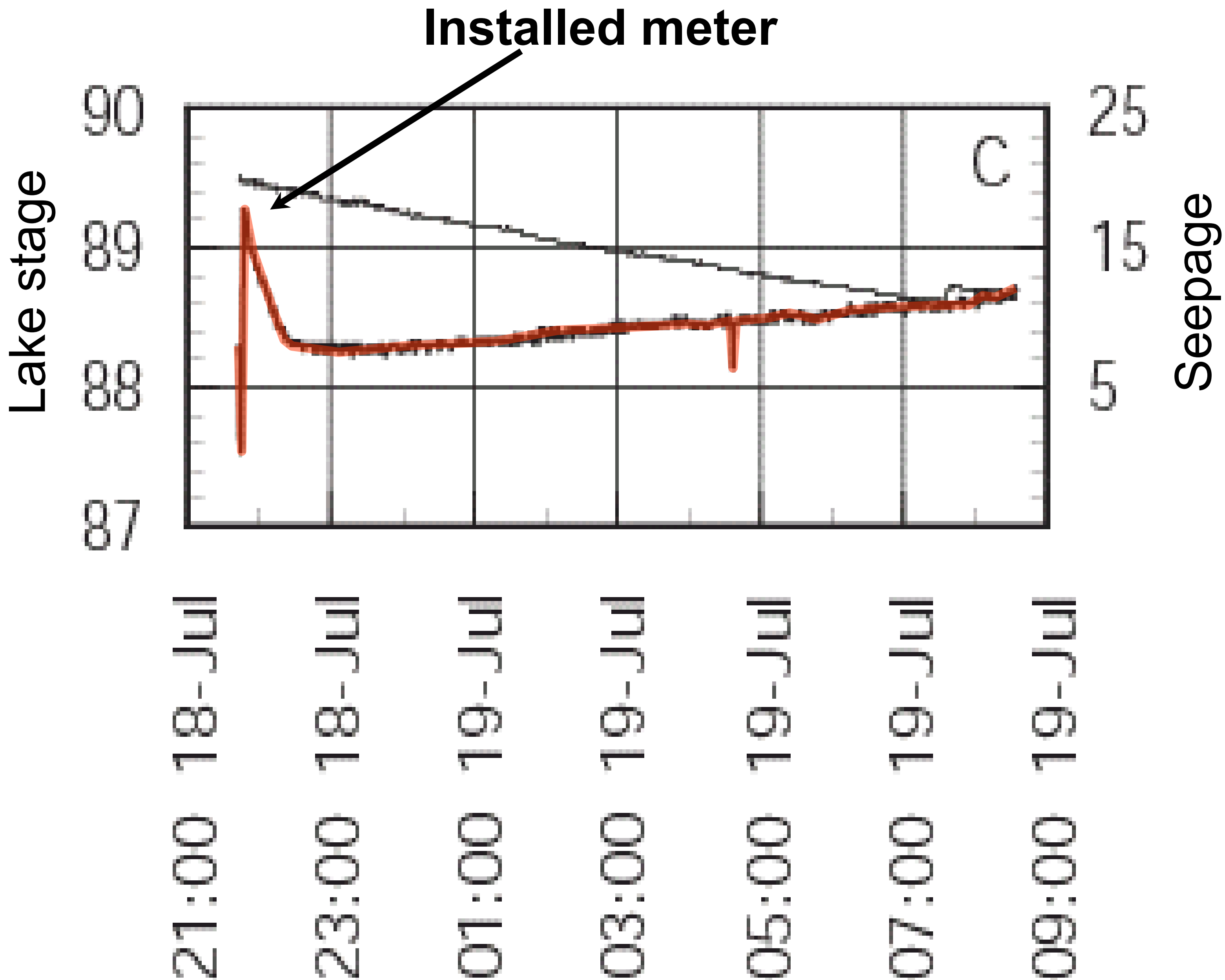
This is the same place where you saw data from many seepage meters installed in a small area. Remember the slide that showed seepage ranging from about 5 to 150 cm/day?

Rosenberry & Morin, 2004

These data are amazing and would not have been collected if we had used manual seepage meters. Note the time of day! I was asleep when these data were collected. Here's what happened. A thunderstorm arrived on the scene at 00:45. Rapid seepage was downward before the storm and was flowing at about -138 cm/day. Wind at the beginning of the storm pushed lakewater to the other side of the lake, reducing lake stage and, therefore, the hydraulic gradient. Seepage decreased to about -115 cm/day as a result. Rain began to fall and the lake stage rose rapidly about 5 mm. Seepage increased to almost -150 cm/day. Then lake stage kept rising, although more slowly, but seepage began decreasing. This was because the rain on the land next to the lake was now infiltrating to the water table. This reduced the hydraulic gradient, which reduced seepage. Pretty cool, huh? And I slept right through the thunderstorm and didn't even know that it had rained until I saw these data.




Rosenberry & Morin, 2004

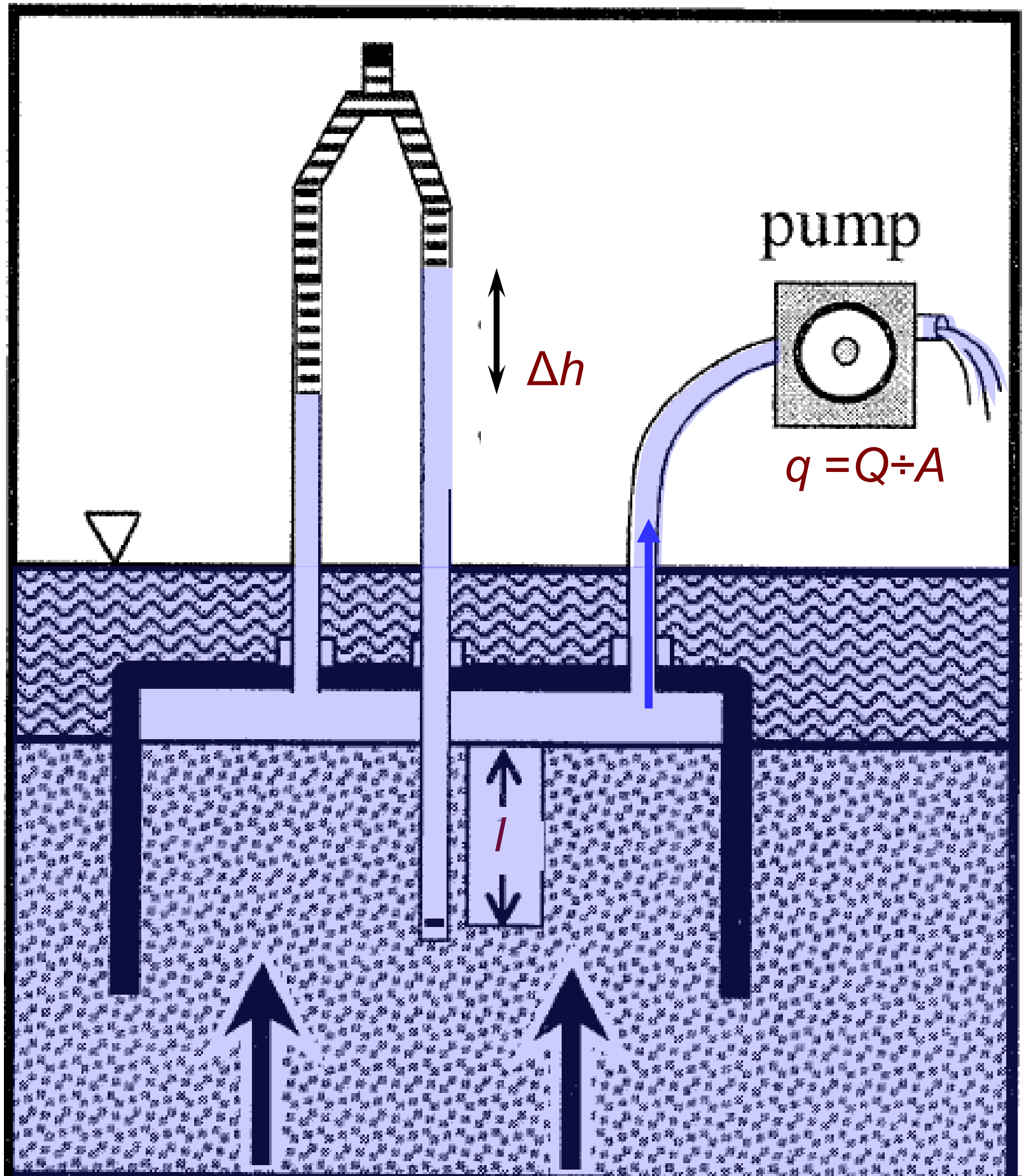



Here’s a dataset that gives us information about how long it takes for disturbance to dissipate before making our first bag measurement. These data indicate we should wait about an hour for the disturbance to dissipate.

Rosenberry & Morin, 2004

# Piezoseep

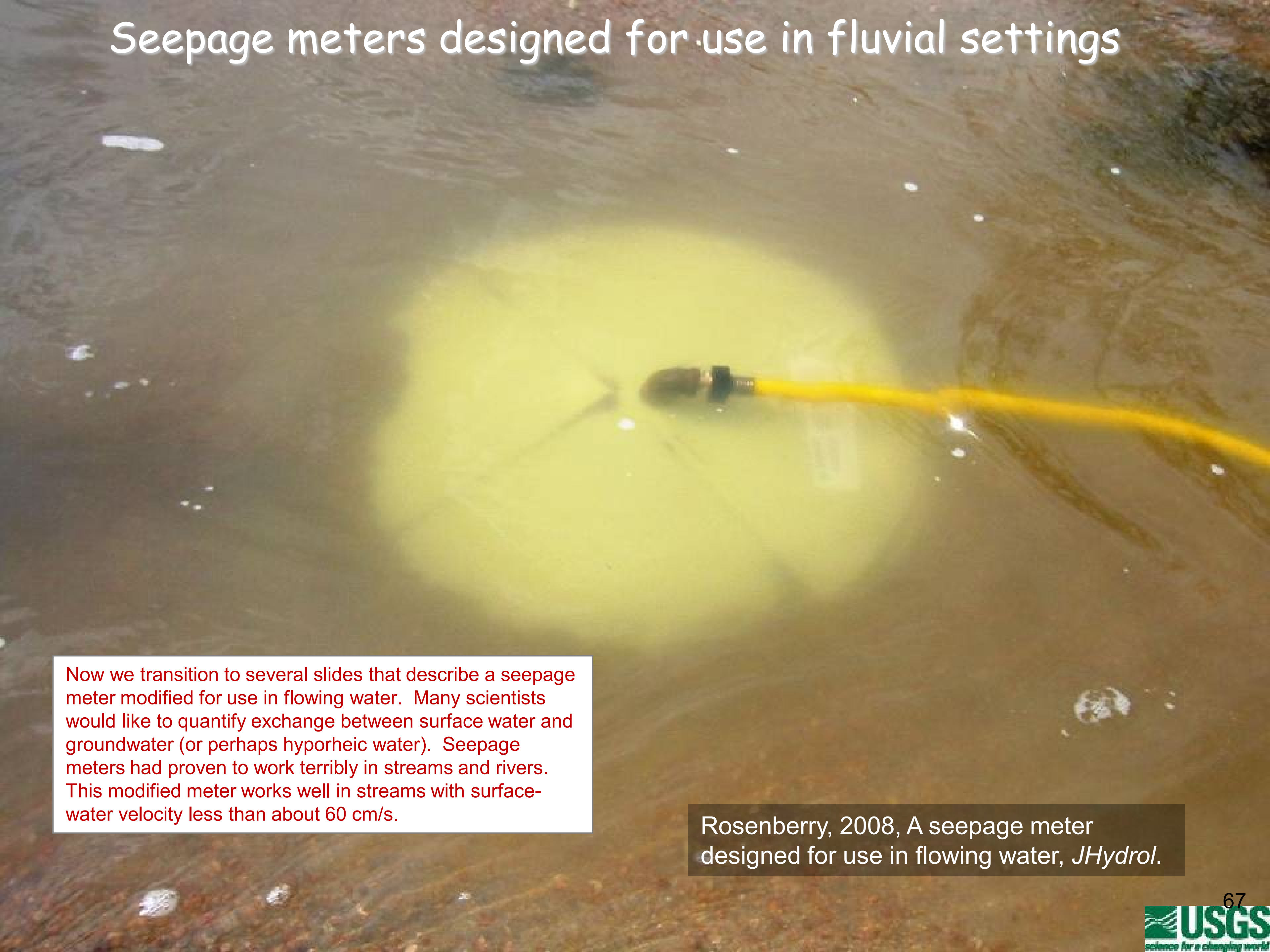
- Combined a piezometer and a seepage meter
- Relate seepage rate to head differential for a range of pumping rates
- Once calibrated, simply measure seepage by accurately measuring Δ*h*




Murdoch, L.C., and S.E. Kelly. 2003. Factors affecting the performance of conventional seepage meters. Wat. Resour. Res. **39**:doi:10.1029/2002WR001347.

This is a really clever modification of a seepage meter. They install the meter that includes a piezometer that extends beneath the seepage cylinder. They pump water from the seepage cylinder (simulating seepage) and record the response in head difference. Once the meter is calibrated to a range of pumping rates, then then simply turn off the pump, log the head difference with a pressure transducer, and then relate that head difference to seepage. No seepage bag!

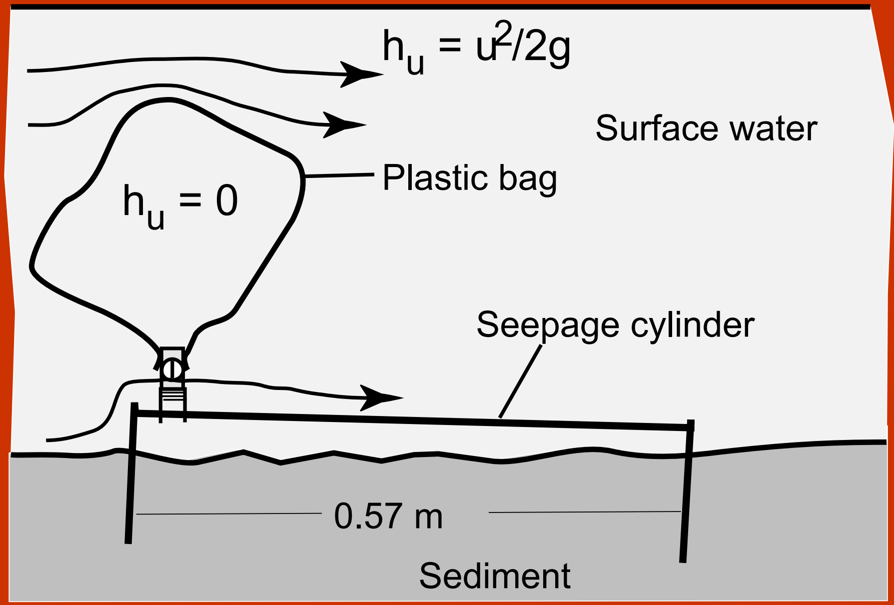
# Seepage meters designed for use in fluvial settings



Now we transition to several slides that describe a seepage meter modified for use in flowing water. Many scientists would like to quantify exchange between surface water and groundwater (or perhaps hyporheic water). Seepage meters had proven to work terribly in streams and rivers. This modified meter works well in streams with surface-water velocity less than about 60 cm/s.

Rosenberry, 2008, A seepage meter designed for use in flowing water, *JHydrol*.

$h_{in} = h_{out} - u^2/2g$




## "Seepage meters and Bernoulli's revenge"

Shinn et al., 2002, *Estuaries*

- Asbury, 1990
- Libelo & MacIntyre, 1994
- Schneider, 1994
- Cable et al., 1997
- Sebestyen & Schneider, 2001

- "Comment on Bernoulli's revenge" – Corbett & Cable, 2003, *Estuaries*
- "Exonerating Bernoulli?" – Cable et al., 2006, *L&O*


The basic problem is due to velocity head, something that hydrogeologists rarely think about because it generally can be ignored in groundwater flow. In this case, a head gradient is generated across a seepage bag with a flexible membrane that causes the bag to fill with water when a current is passing by the bag. A velocity-generated pressure gradient also exists across the seepage cylinder, but it is a rigid device and can withstand the pressure gradient with no deformation. But the gradient can cause water to flow beneath the edge of the meter.

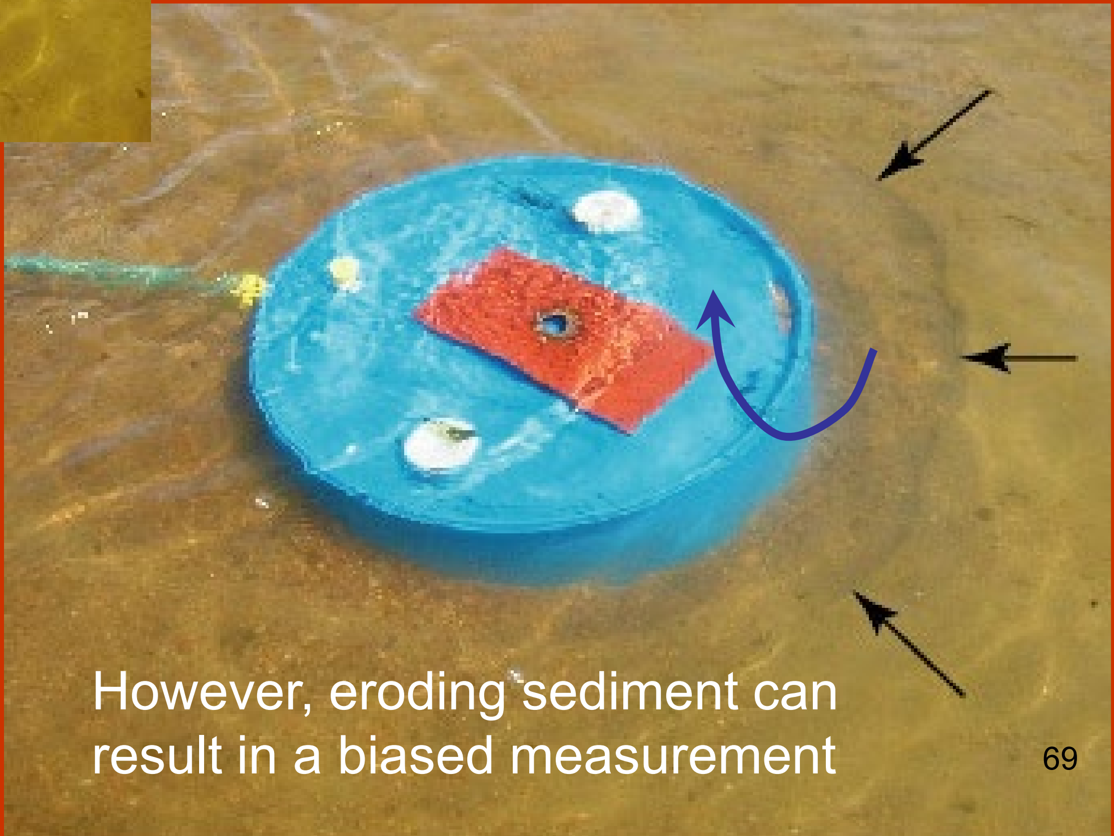
But what about the seepage cylinder?

Offsetting influences, and a rigid cylinder, usually result in minimal net influence.

Fortunately, the low-velocity, high pressure areas (red) are balanced by high-velocity, low pressure areas (green), so the net effect on seepage around the cylinder is minimal. This statement is based on controlled experiments in a seepage tank.

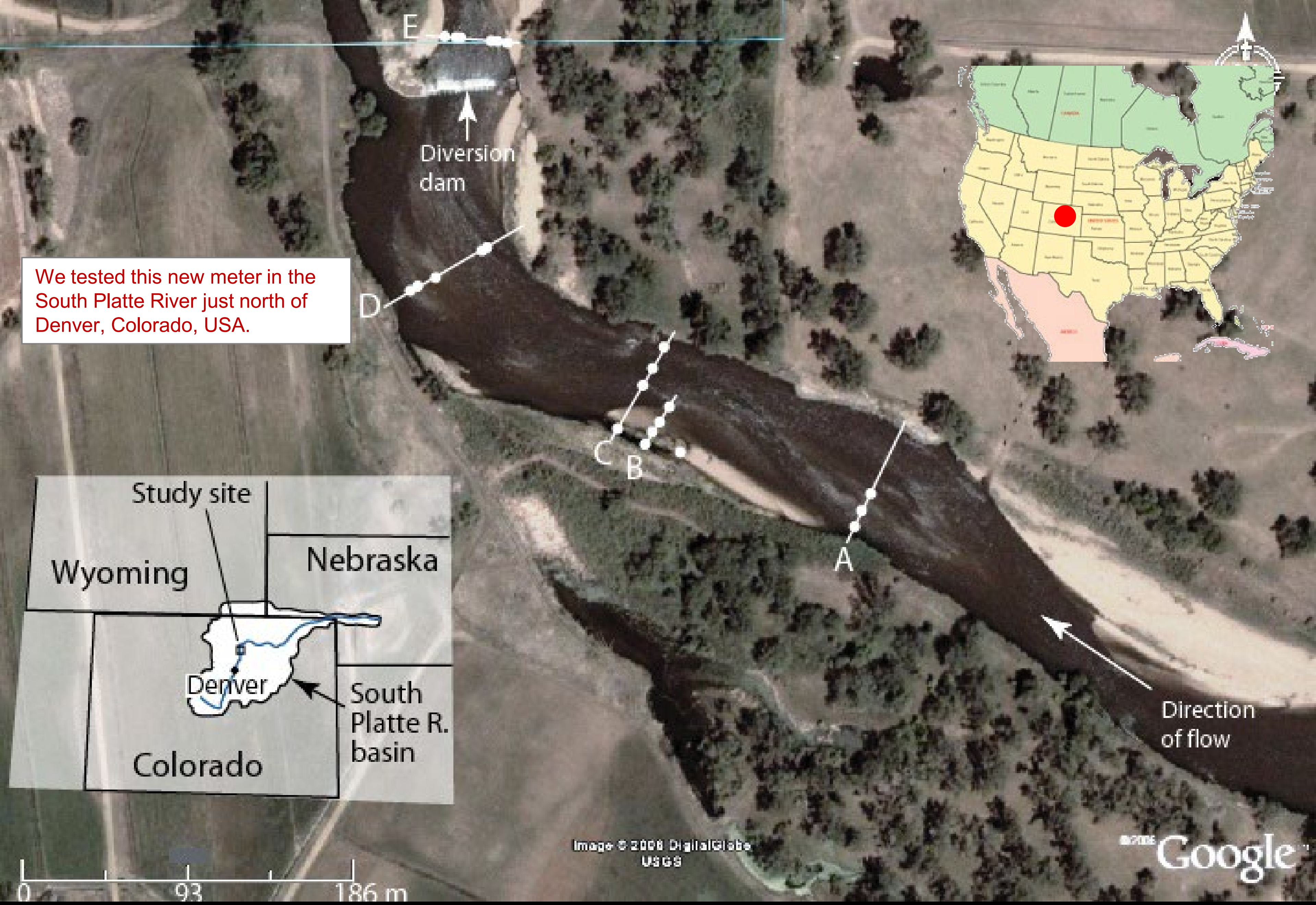
Rosenberry & Pitlick, 2009, *JHydrol*.

However, remember that substantial errors can result in high-*K* settings, as we saw a few slides ago.



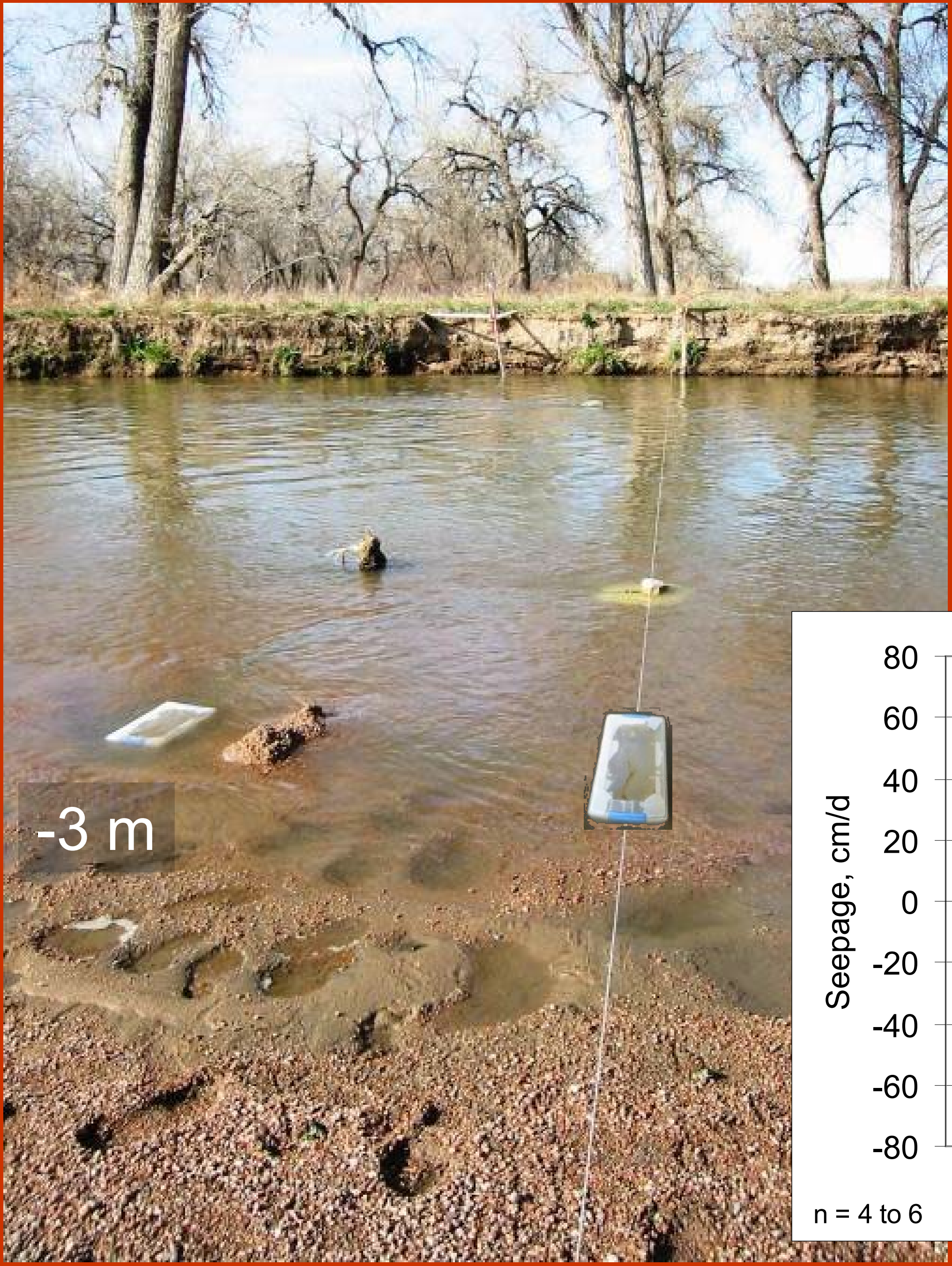
# Modified seepage meter

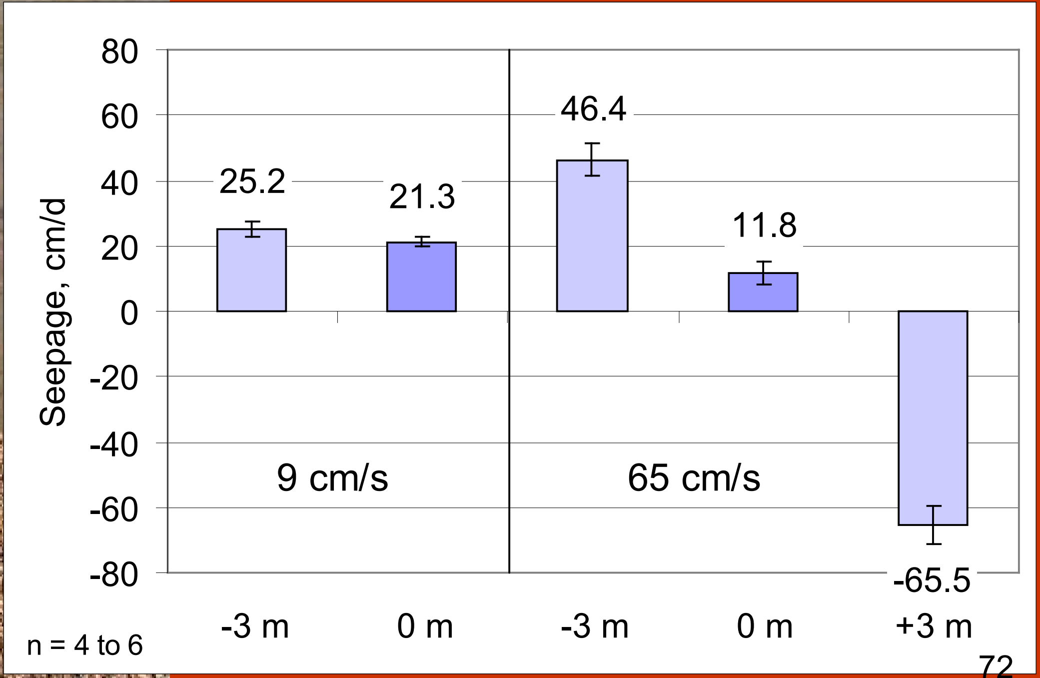
Here we use a low-profile seepage cylinder that is less likely to be pushed out of the bed by the current, and a bag shelter that is placed where velocity is not quite as fast.


E
Diversion dam
We tested this new meter in the South Platte River just north of Denver, Colorado, USA.
D
C
B
A
Direction of flow
Study site
Wyoming
Nebraska
Denver
South Platte R. basin
Colorado
Image © 2008 DigitalGlobe
USGS
Google
0
93
186 m

# Bag placement can be <u>very</u> important

If making measurements in a fast-moving stream you need to make sure you place the bag shelter at the same river elevation as the seepage cylinder. If you place the bag downstream, the bag will fill up. If you place the bag upstream, it will lose extra water. The data below indicate that bag placement was not a concern when river velocity was 9 cm/s, but it was a big concern when river velocity was 65 cm/s.

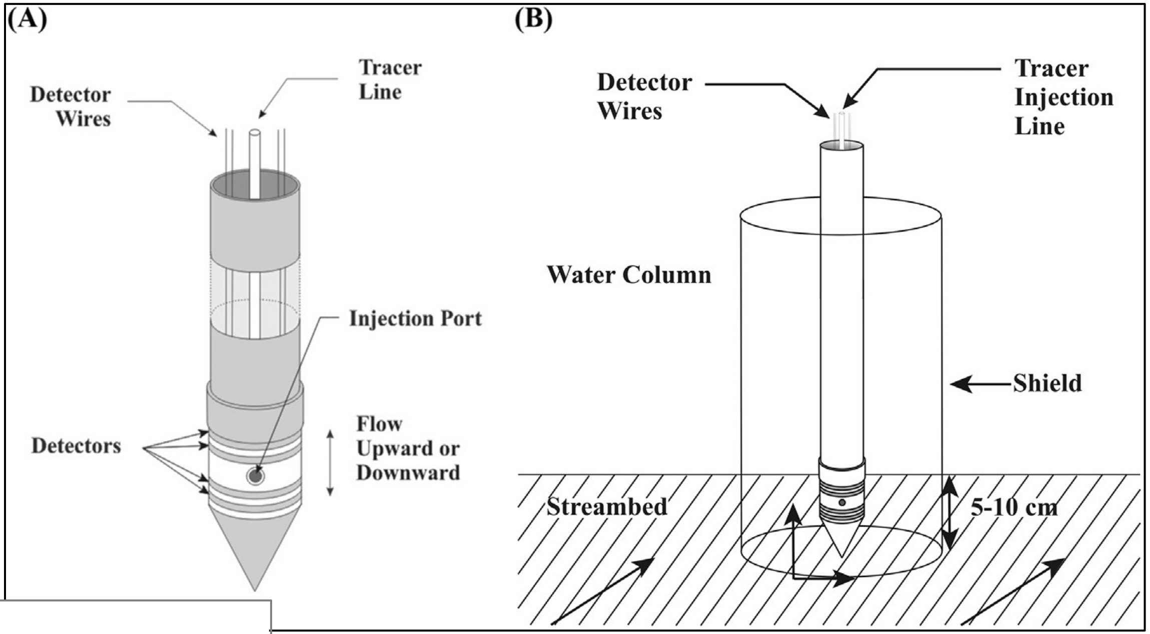


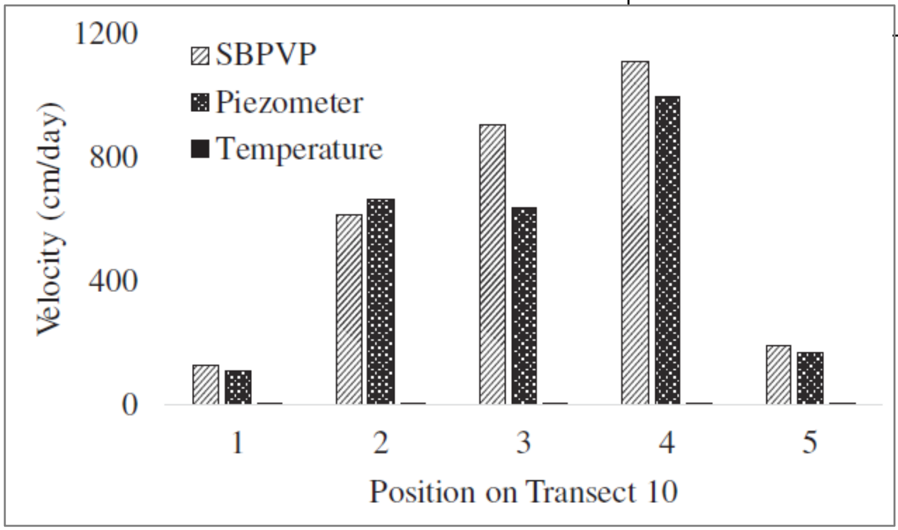



# Other methods: Point-velocity probe (tracer-injection)

The streambed point-velocity probe releases a salt tracer and then measures input from salinity detectors above and below the point of tracer release. Salinity detectors are monitored each second by a digital datalogger. Timing of the breakthrough curve is related to seepage velocity by fitting the curve to a 1-D advective-dispersion equation.

An open-ended cylinder is installed around the probe to eliminate horizontal components of flow that are common in hyporheic settings. The device produced data similar to that obtained from a mini piezometer. Seepage rates from vertical-temperature profiling data were an order of magnitude smaller. The authors state that was likely caused by flow that was primarily horizontal rather than vertical.



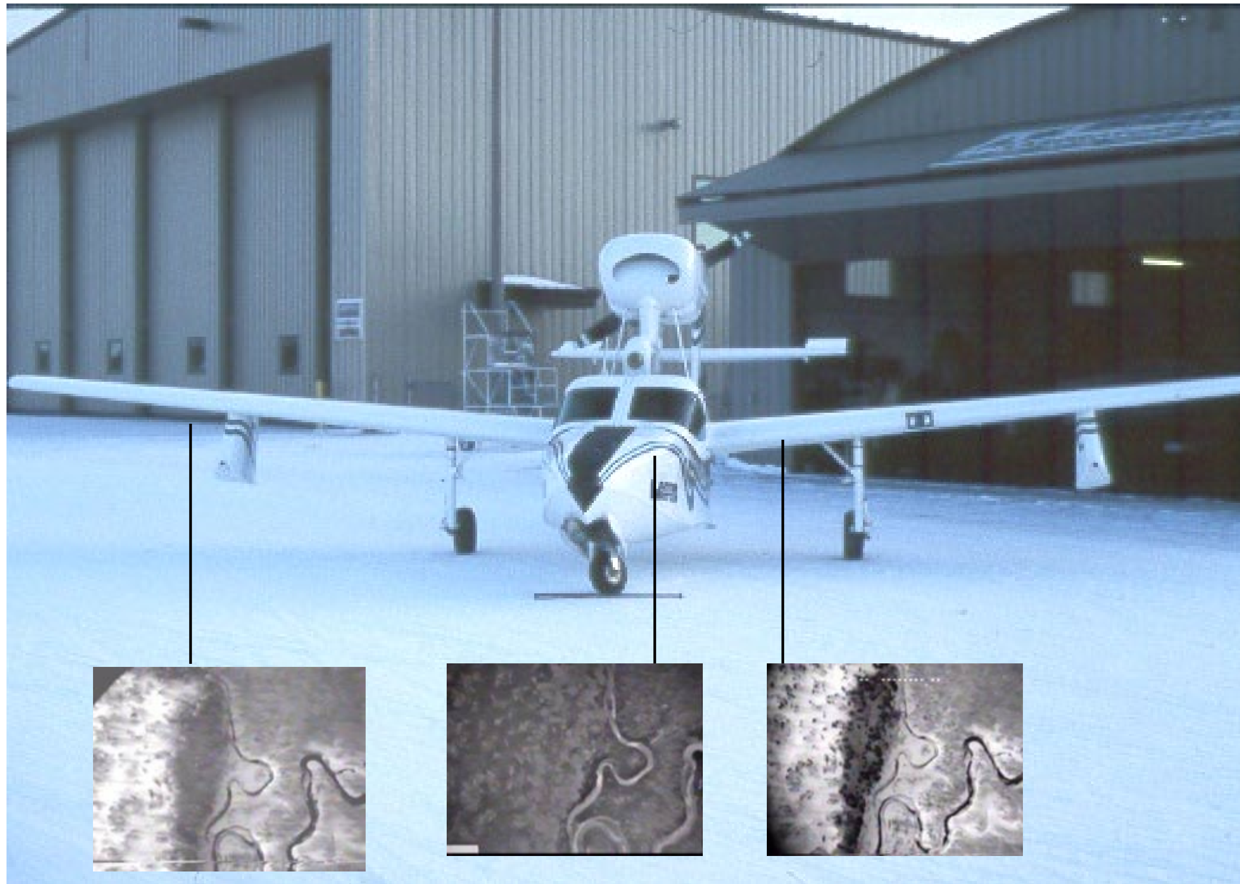

Range of measurable flow is 50 to 450 cm/d.

Cremeans & Devlin, 2017, *JContHydrol*.
Cremeans et al., 2020, *GWMR*

Methods-comparison paper

**Replace these slides with more modern studies of drone-based imagery and results**

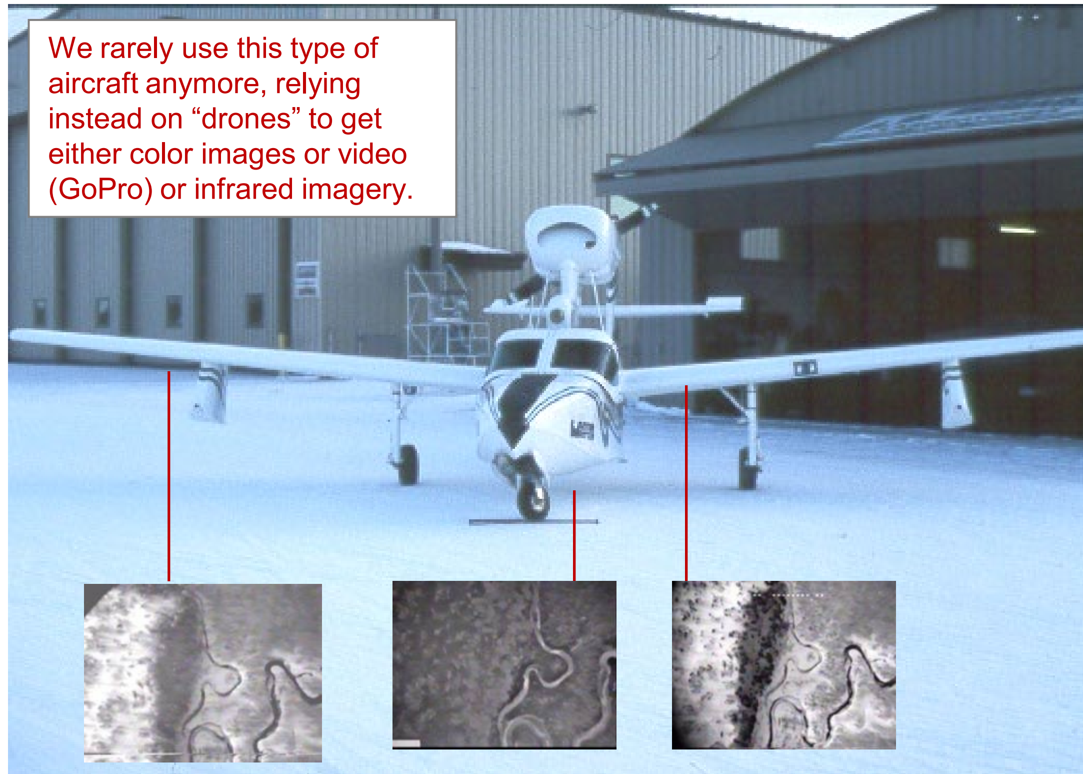
# Qualitative approach: Aerial imagery



*AW Research Laboratories*

We can also use the difference between the temperature of groundwater that discharges to surface water, and the temperature of the surface water, as a qualitative indicator of where groundwater discharge is focused. We will now look at several examples starting with aerial imagery or photographs taken from the air.

# Qualitative approach: Aerial imagery

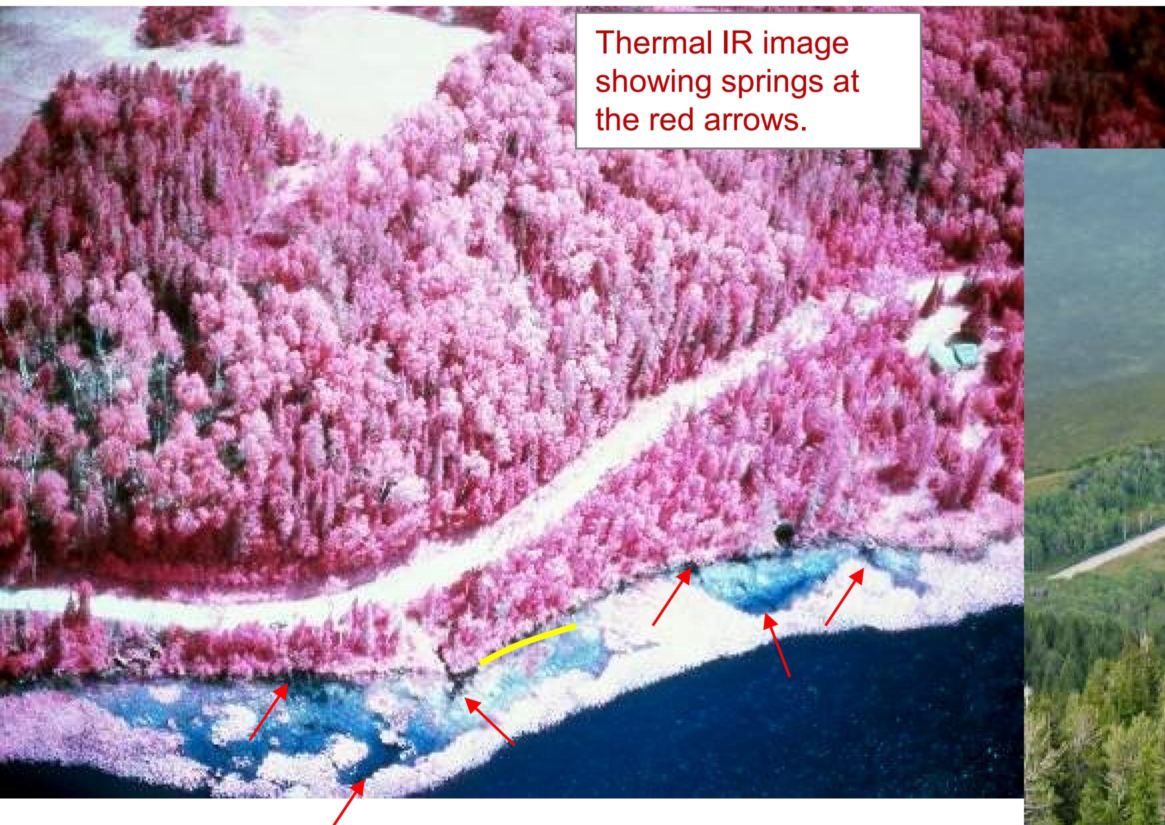



*AW Research Laboratories*

# Aerial imagery

- Usually thermal infra-red (IR)
- Temperature difference between SW and GW needs to be greater than sensitivity of the film
- Springs during ice cover are easy to spot
- Sometimes GW discharge can be seen with visible spectrum imagery

Open water on the edge of an ice-covered lake



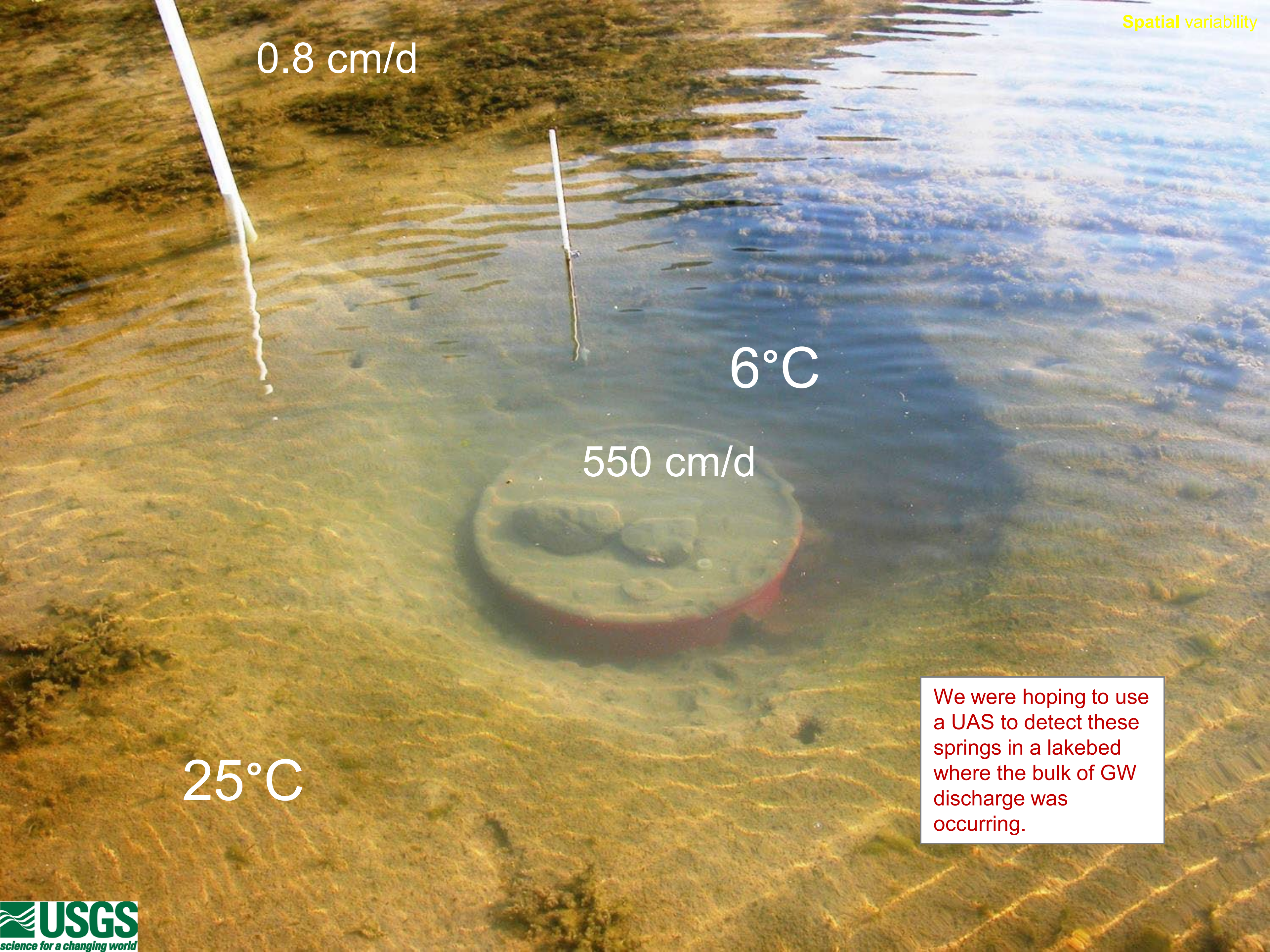
Thermal IR image showing springs at the red arrows.


Here you can see springs with a regular camera if you know what to look for.

# USGS “Air Force”


This was military grade, cutting-edge technology in the early 1990s that the Army gave to USGS, but it’s old-school now.

Spatial variability

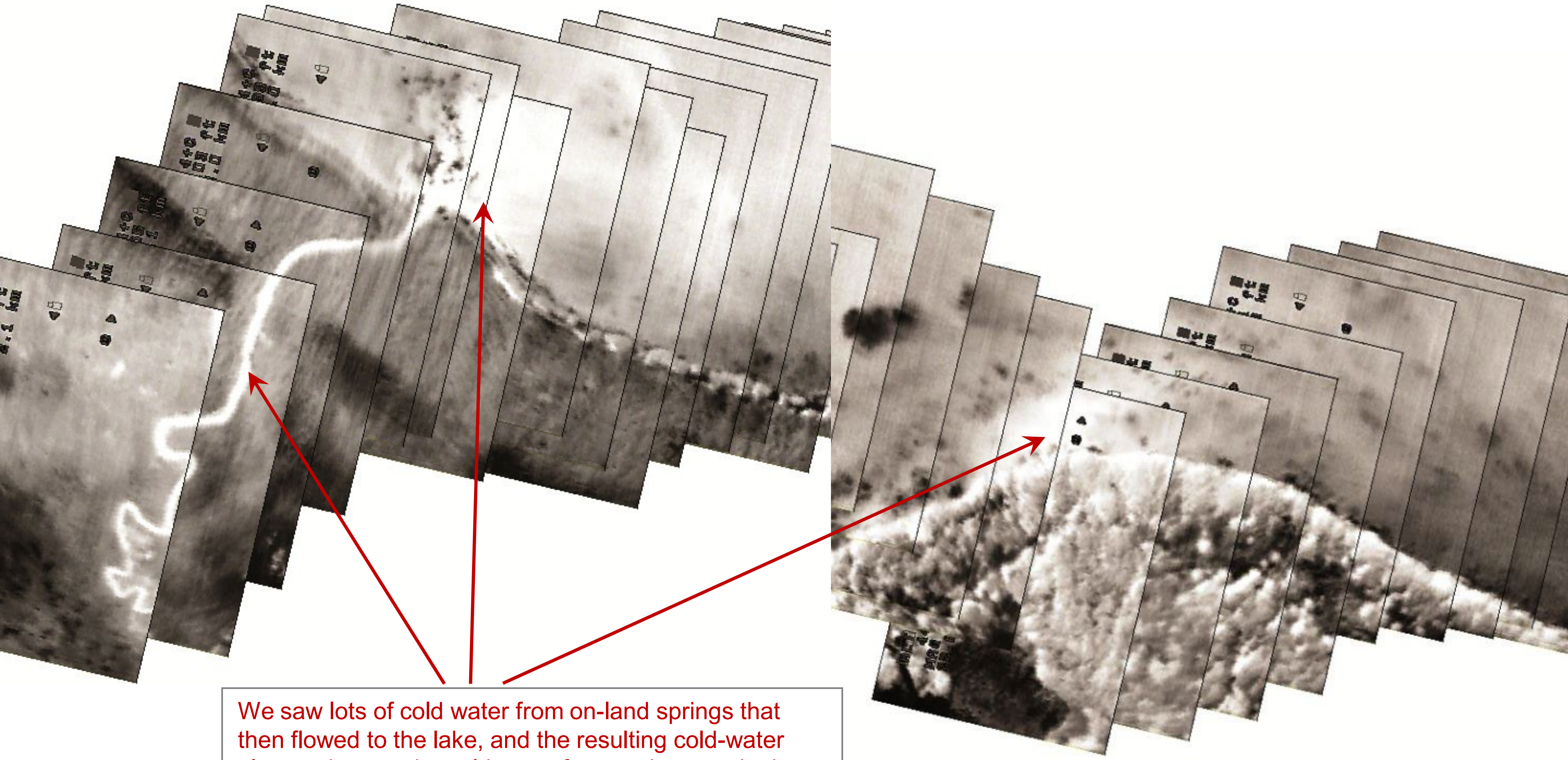



We were hoping to use a UAS to detect these springs in a lakebed where the bulk of GW discharge was occurring.

USGS science for a changing world


We flew on a nice,
calm day

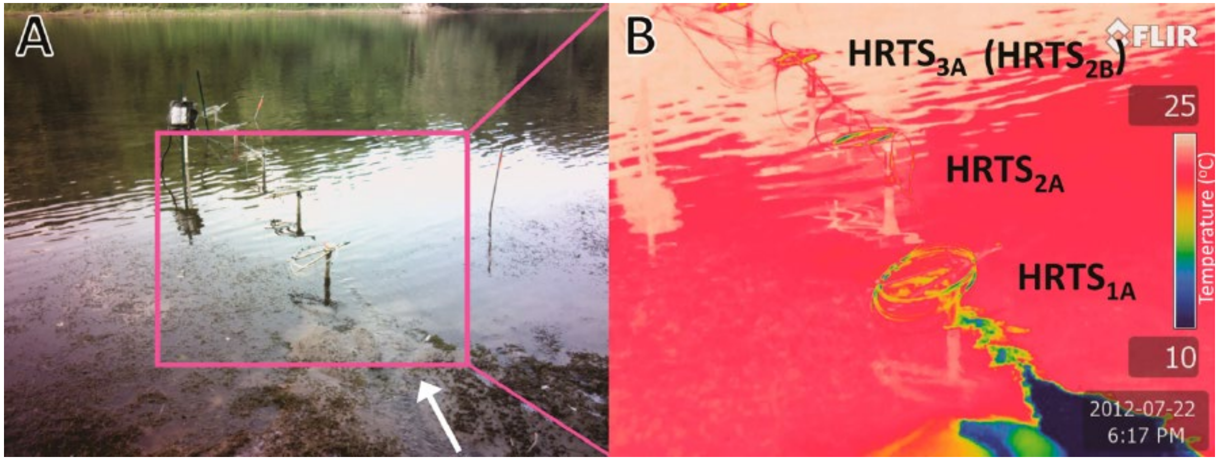
And we couldn't see these large springs in the imagery. The cold GW discharge was staying right on the bottom and not mixing to the surface.



We saw lots of cold water from on-land springs that then flowed to the lake, and the resulting cold-water plumes, but not the cold spots from springs out in the lake that we knew were there.

Hare et al., 2015, *JHydrol*.

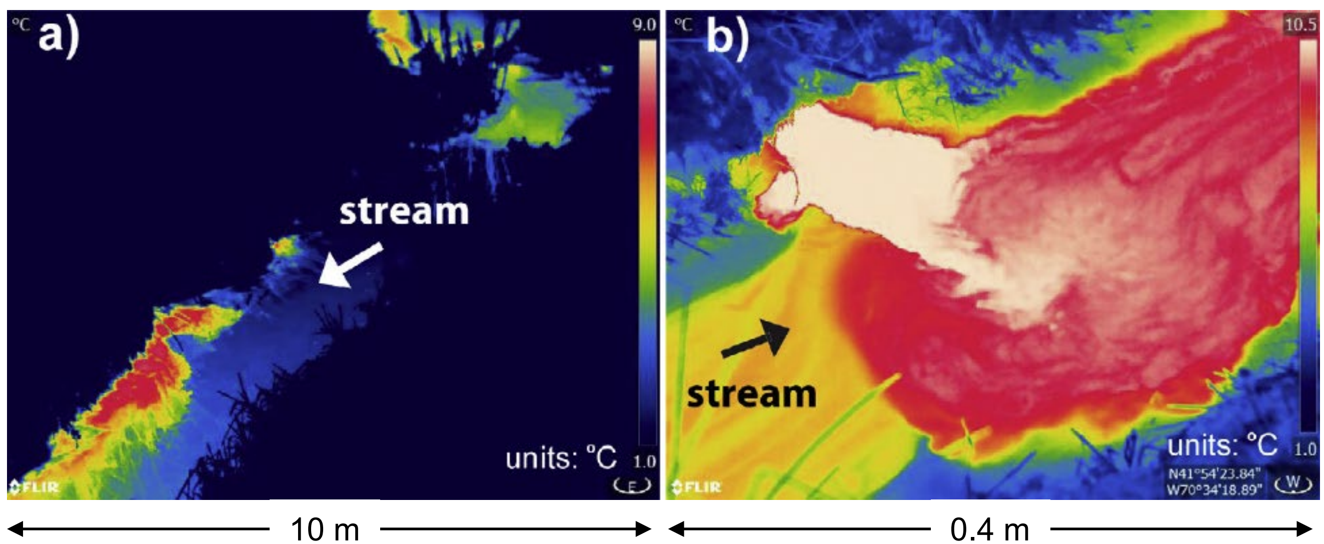
# Hand-held thermal infra-red (TIR)




Briggs et al., 2013, ES&T.

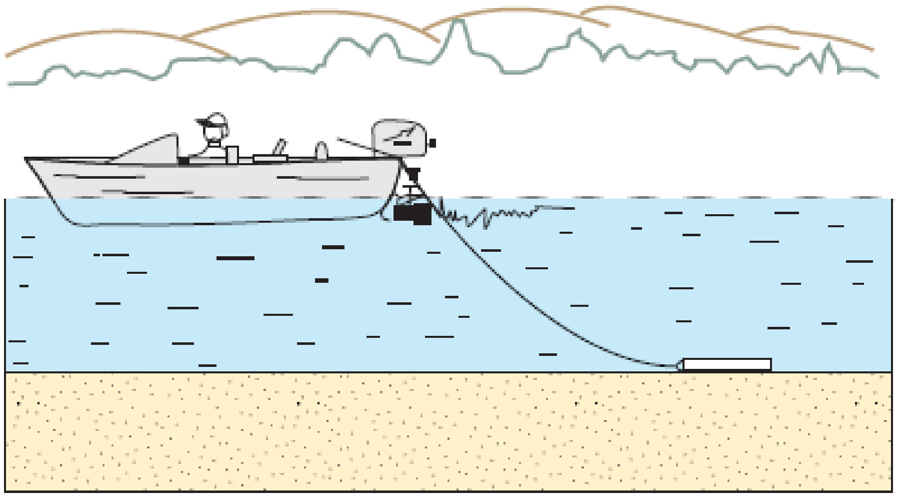
This shows a spring located right at the streambank that discharges cold water to a warmer river during the summer. The cold water plunges beneath the water surface within about 2.3 m from the riverbank. The cold area on the riverbed, however, extended 7 to 8 m from the shoreline.

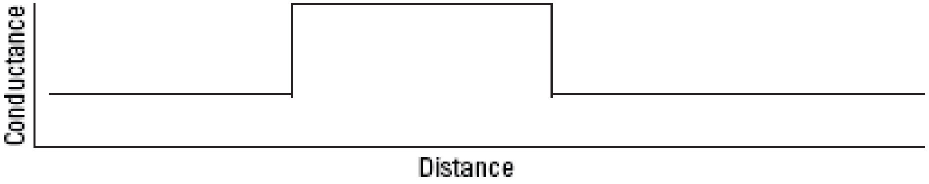
Here we see spring discharge that is warmer than the surface water because the imagery was collected in winter. The warmer groundwater is less dense and moves to the stream surface, where it can be seen with infrared imagery. Image *a* is about 10 m from left to right and image *b* is only 0.4 m from left to right.




Hare et al., 2013, *JHydrol*.

# Towed probes

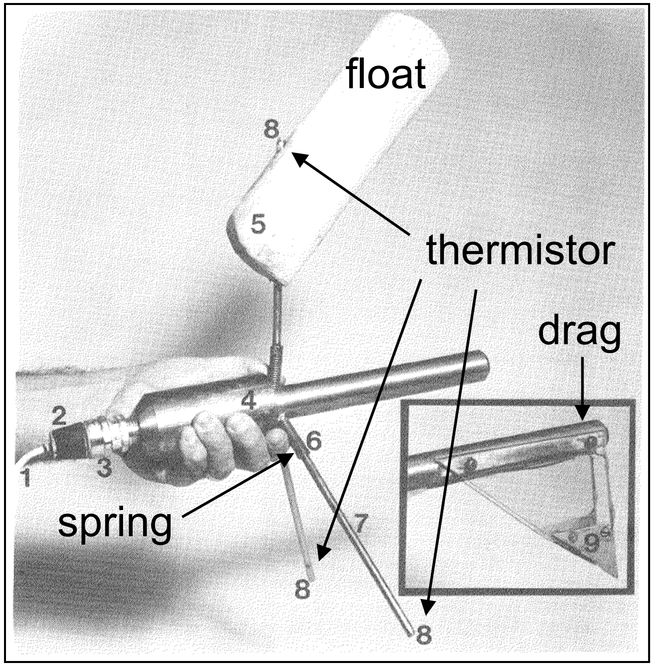





Typically, temperature and specific conductance are the sensors of choice

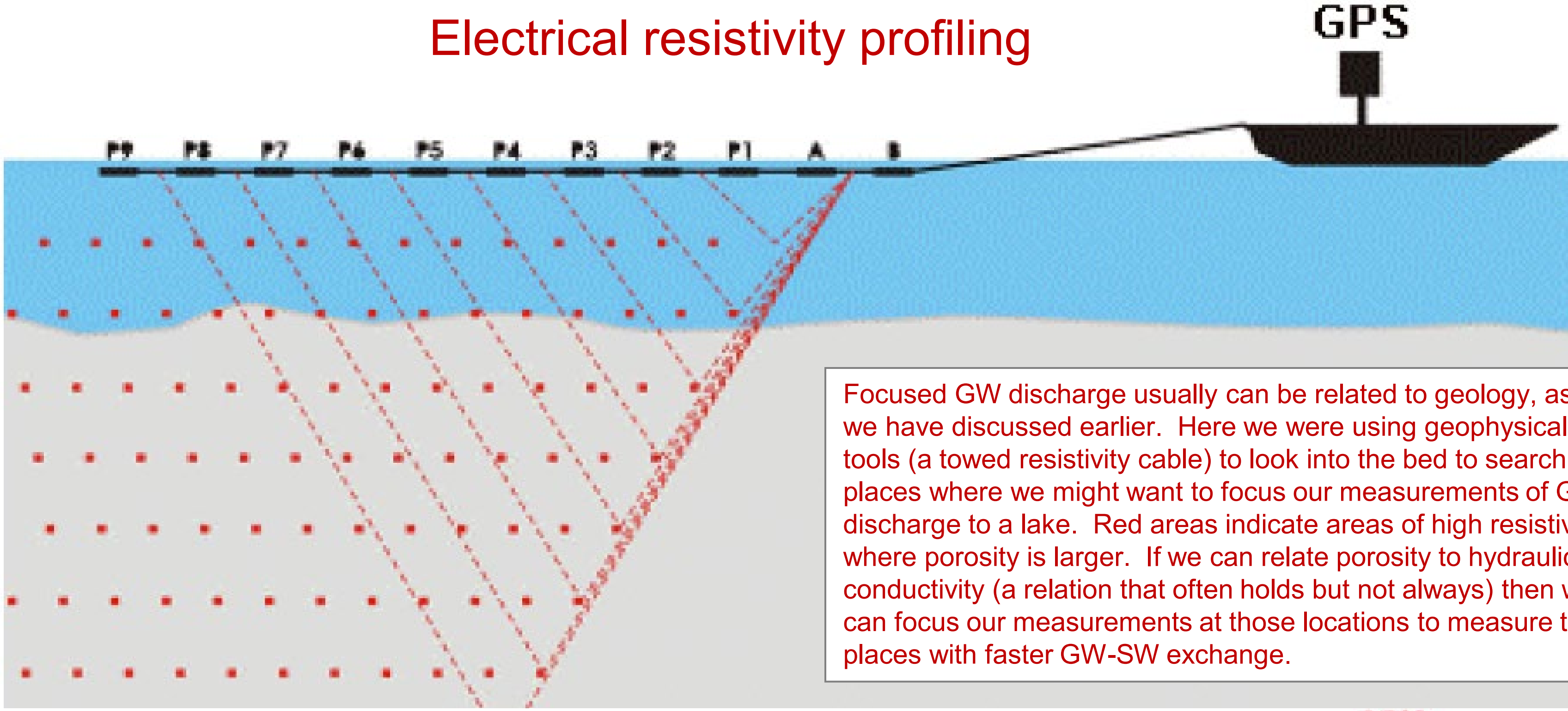
Used in combination with GPS to map a lake or reservoir or river

Remember David Lee, the guy who invented the half-barrel seepage meter? He also invented this clever probe that could be towed behind a boat. It records the temperature and specific conductance of the water right at the bed of a lake. Thermal anomalies can be related to places of focused groundwater discharge.

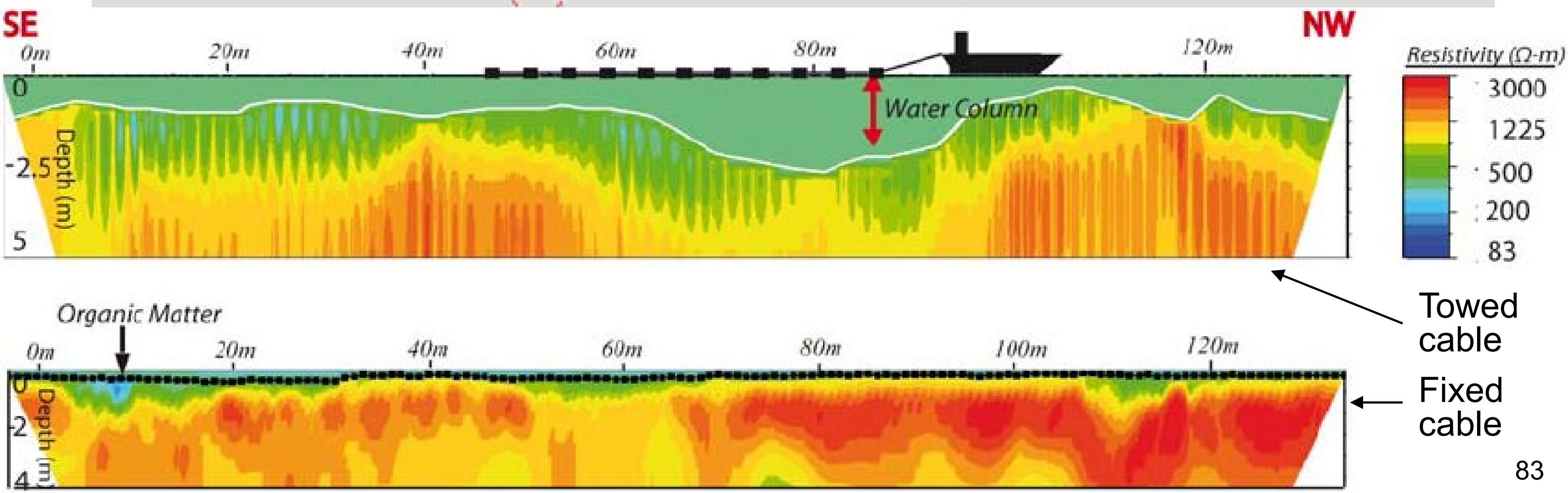



Lee, 1985, *JHydrol.*

# Electrical resistivity profiling




Focused GW discharge usually can be related to geology, as we have discussed earlier. Here we were using geophysical tools (a towed resistivity cable) to look into the bed to search for places where we might want to focus our measurements of GW discharge to a lake. Red areas indicate areas of high resistivity where porosity is larger. If we can relate porosity to hydraulic conductivity (a relation that often holds but not always) then we can focus our measurements at those locations to measure the places with faster GW-SW exchange.



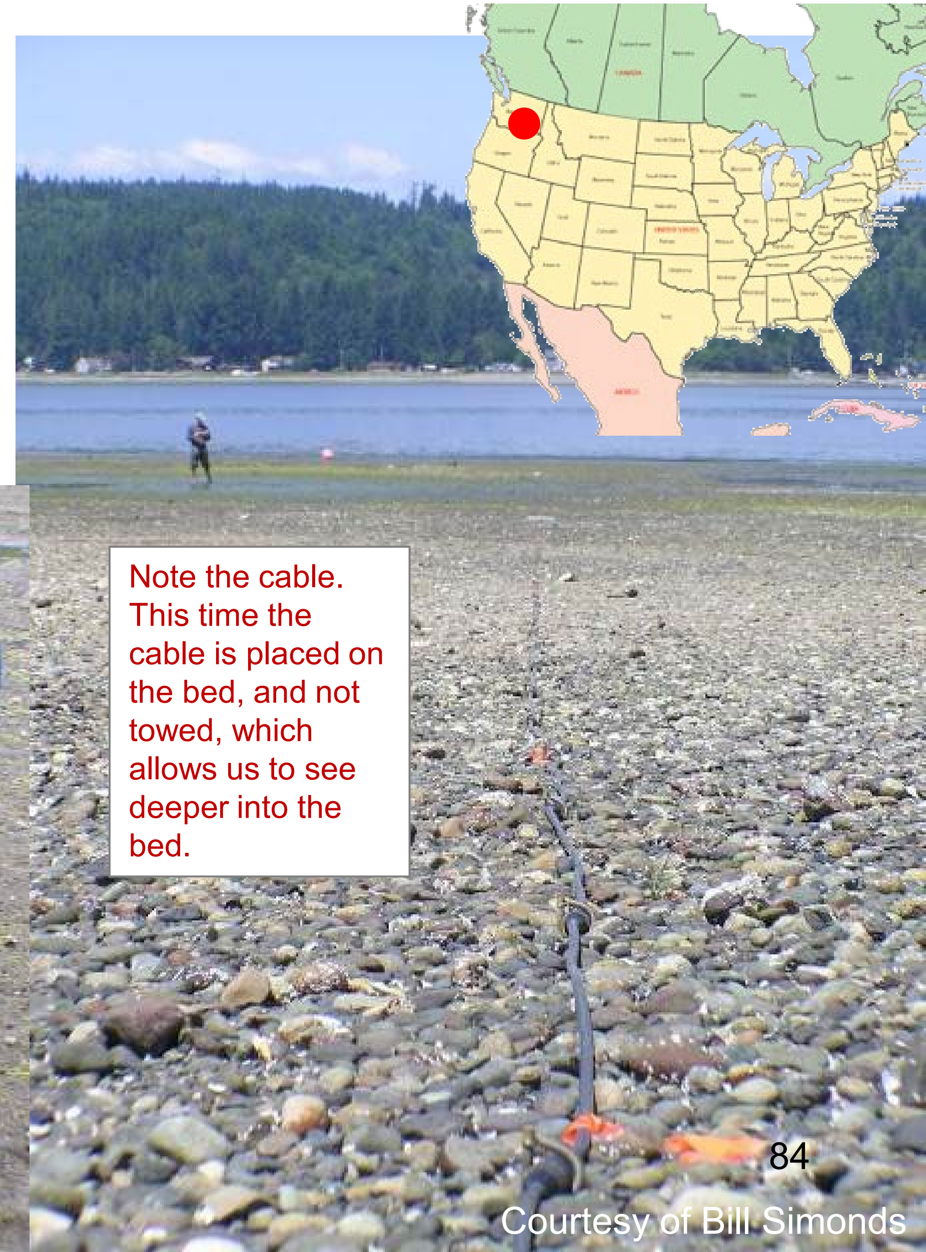

Mitchell et al., 2008, *SAGEEP*

We are using the same method here, but this time the changes in resistivity are caused by salinity. We now are looking for areas of fresh water where GW is discharging to the salty sea. Fresh water is high resistivity (red) and saline surface water has lower resistivity (blue).

# Hood Canal, Puget Sound, WA


Electronics box


Placing the cable on the bed



Note the cable. This time the cable is placed on the bed, and not towed, which allows us to see deeper into the bed.

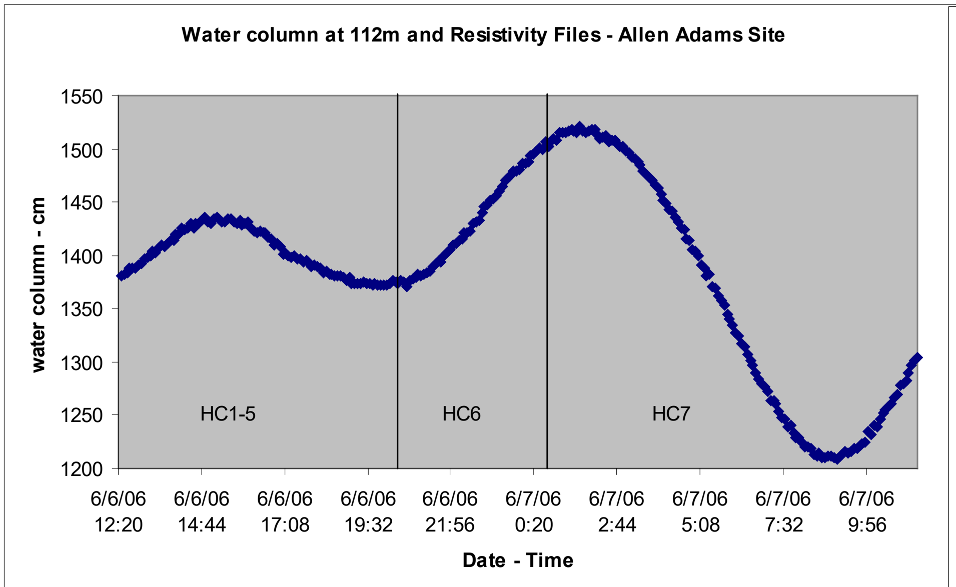
Courtesy of Bill Simonds

# High Resolution Underwater Resistivity

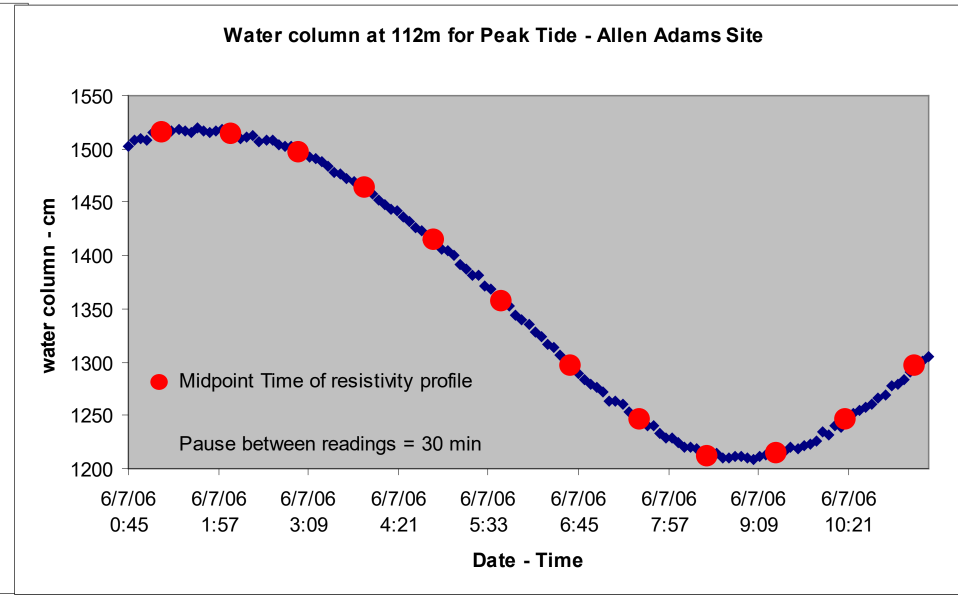
## Hood Canal, WA

Merrimont Time Series

Entire Time Series




Peak Tide Time Series

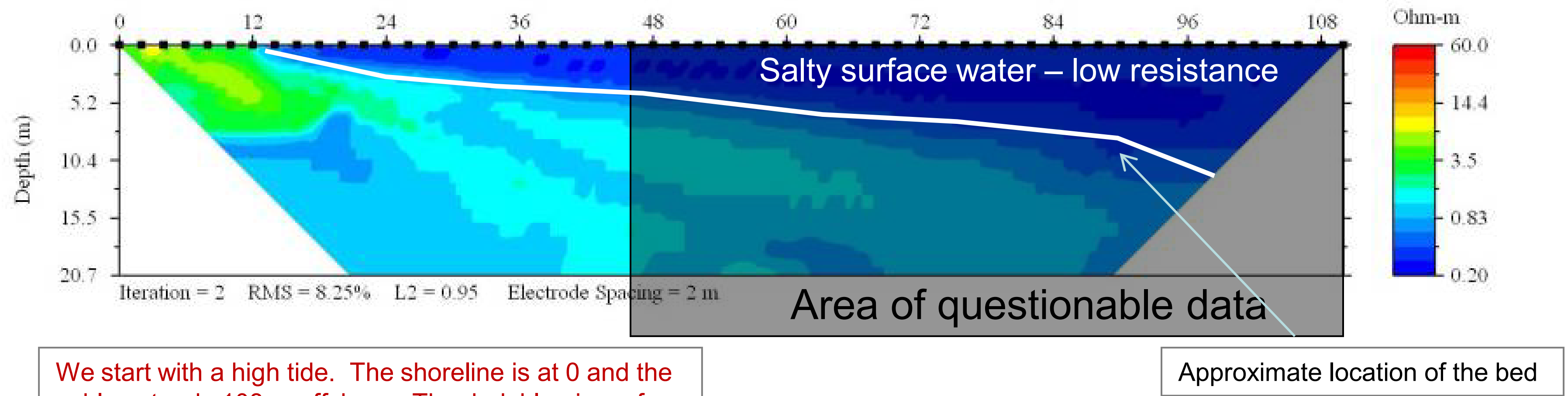



With the cable fixed in one transect, we can also see changes over time. Now you will see a series of slides that show changes in the salt-water fresh-water interface as the tide rises and falls.

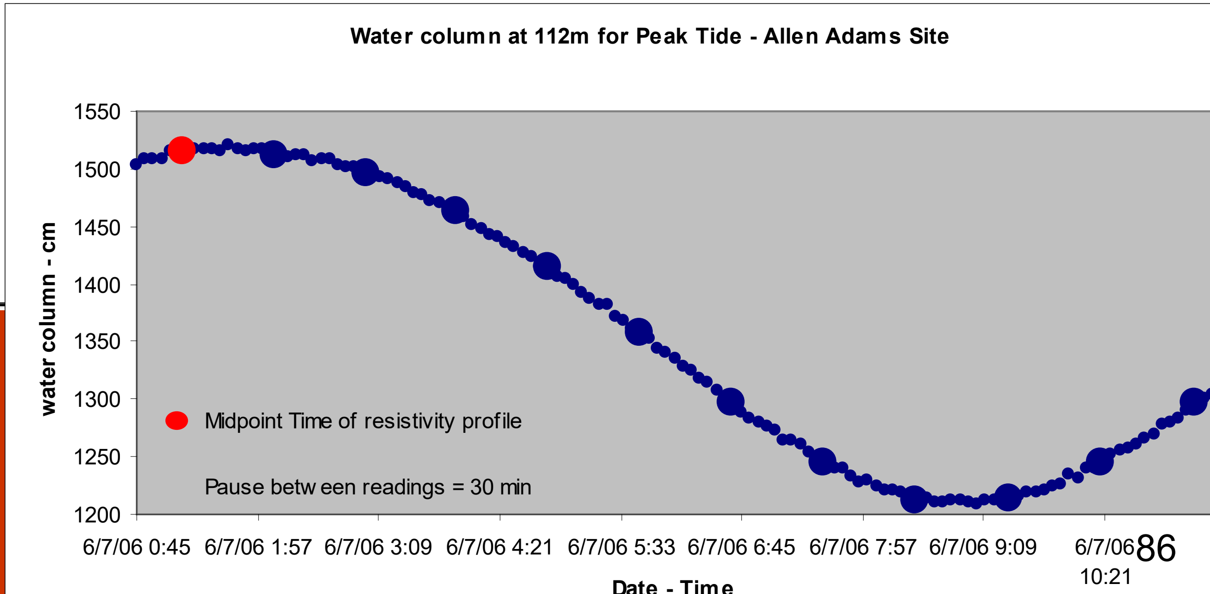


Courtesy of Bill Simonds, USGS

# 1:12



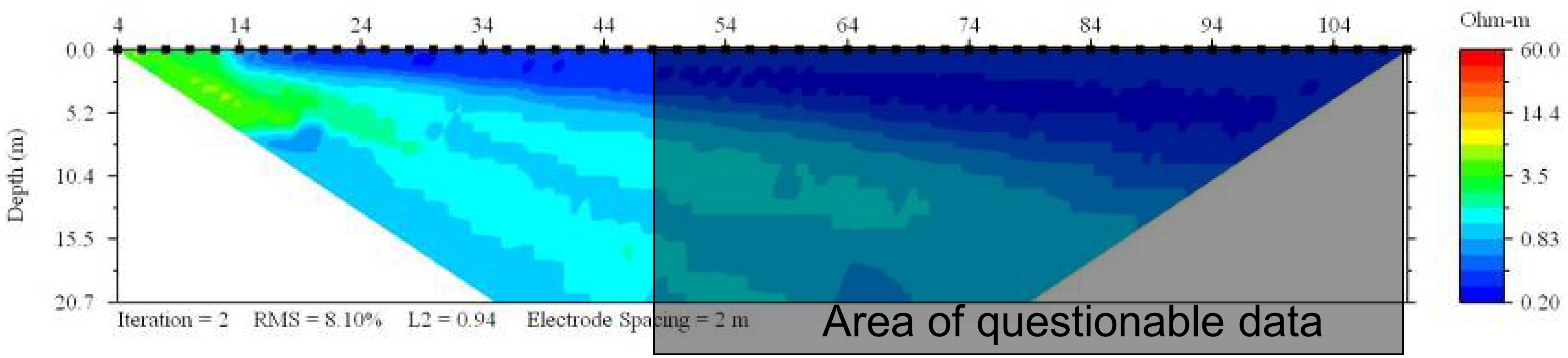

We start with a high tide. The shoreline is at 0 and the cable extends 108 m offshore. The dark blue is surface water. The green and yellow is a mix of GW and saline surface water in the sediments beneath the bed.

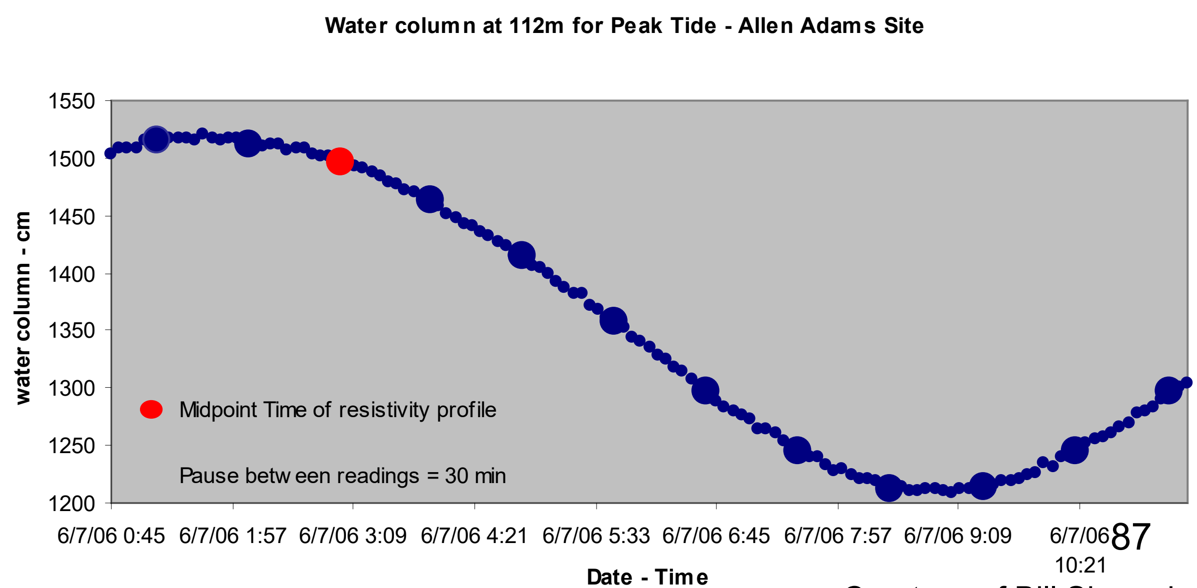





Courtesy of Bill Simonds

3:01

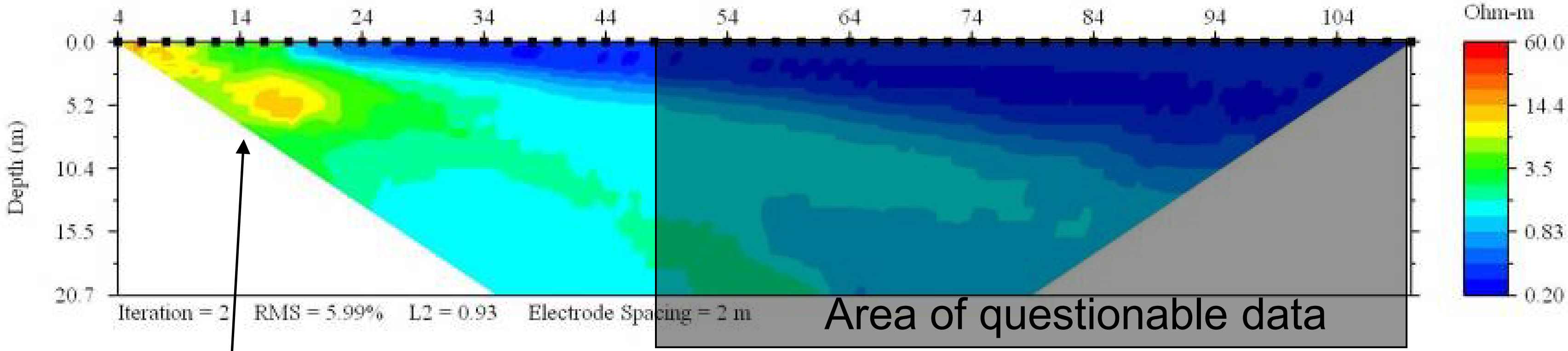




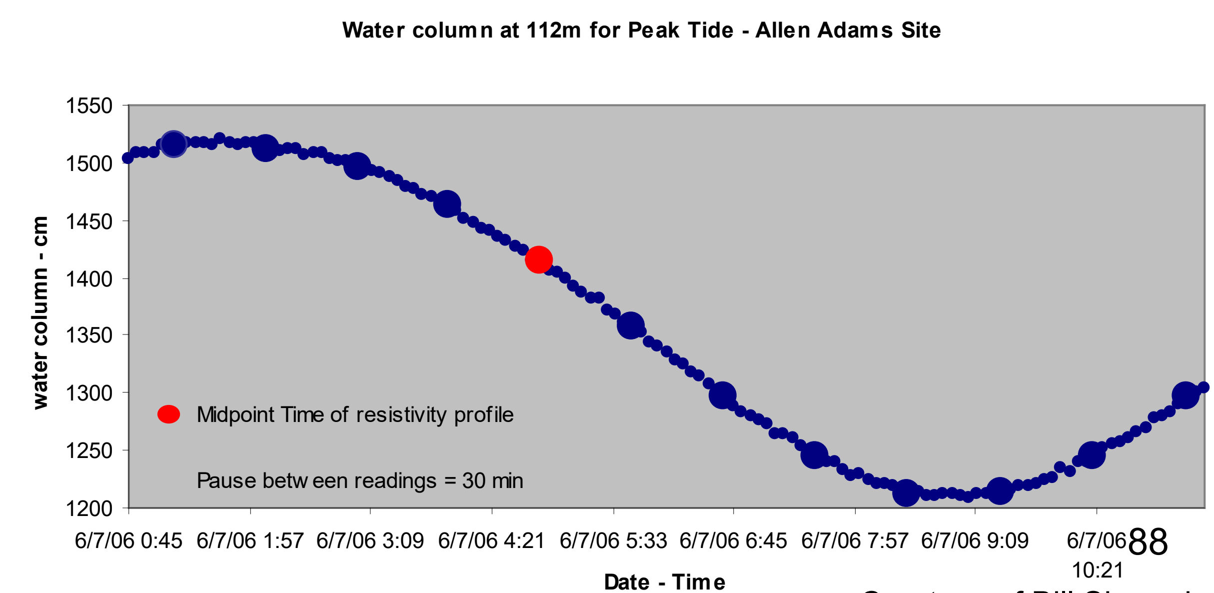

Courtesy of Bill Simonds

# 4:50




Tide going out, fresh-water moving in

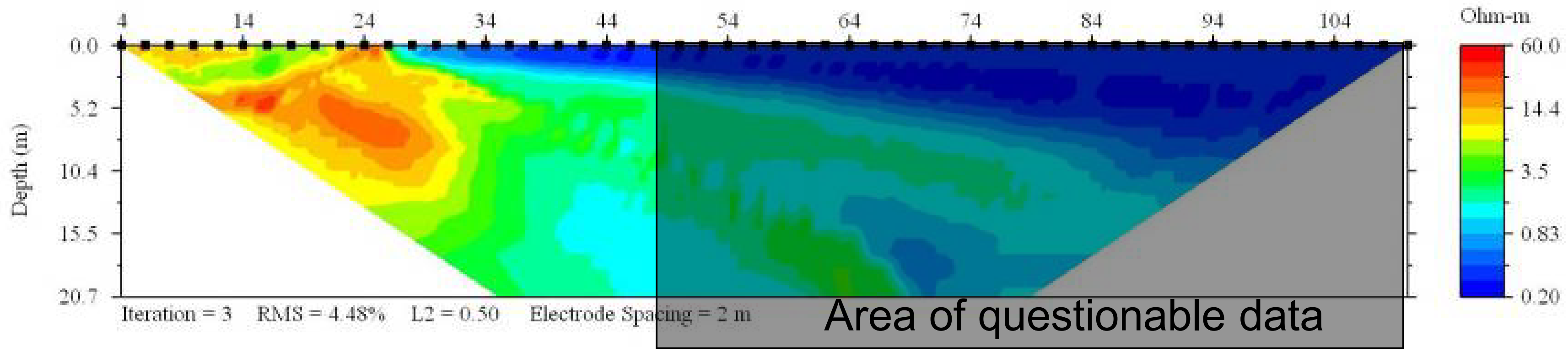
Note that as the tide falls the area of fresher water is extending to the right. Groundwater is now able to discharge to the sea.

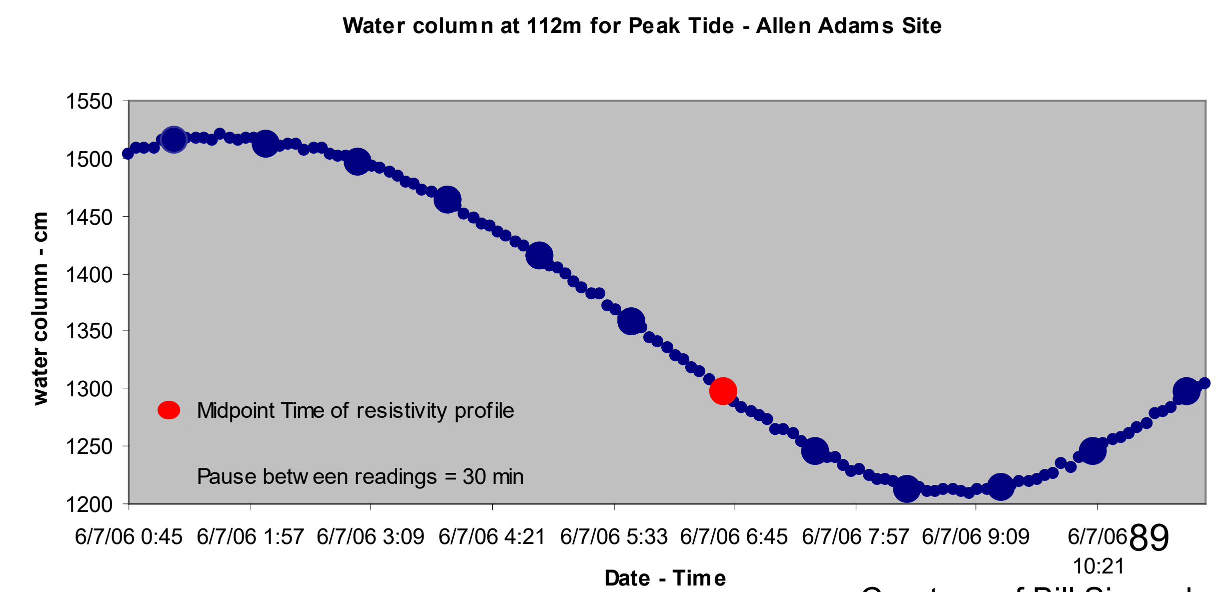





Courtesy of Bill Simonds

6:38

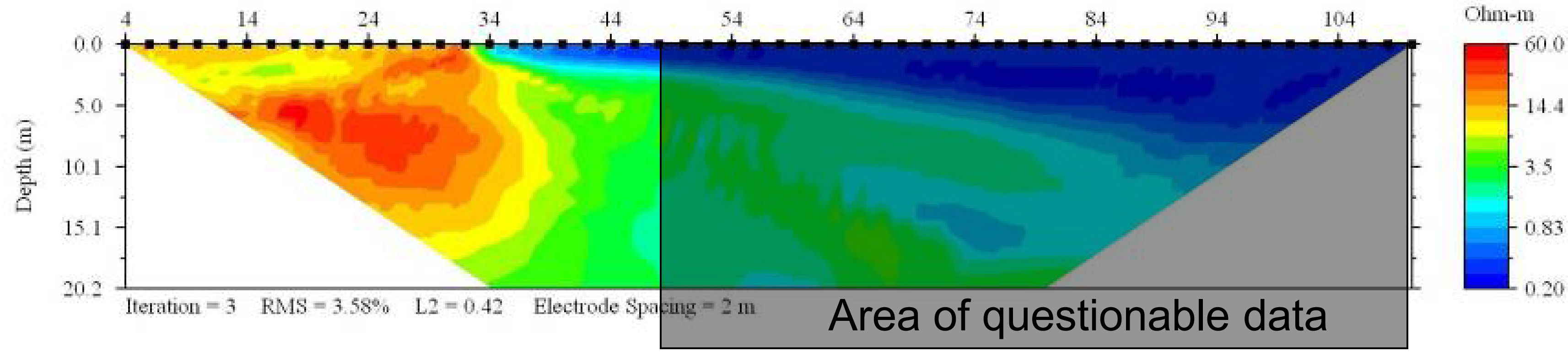


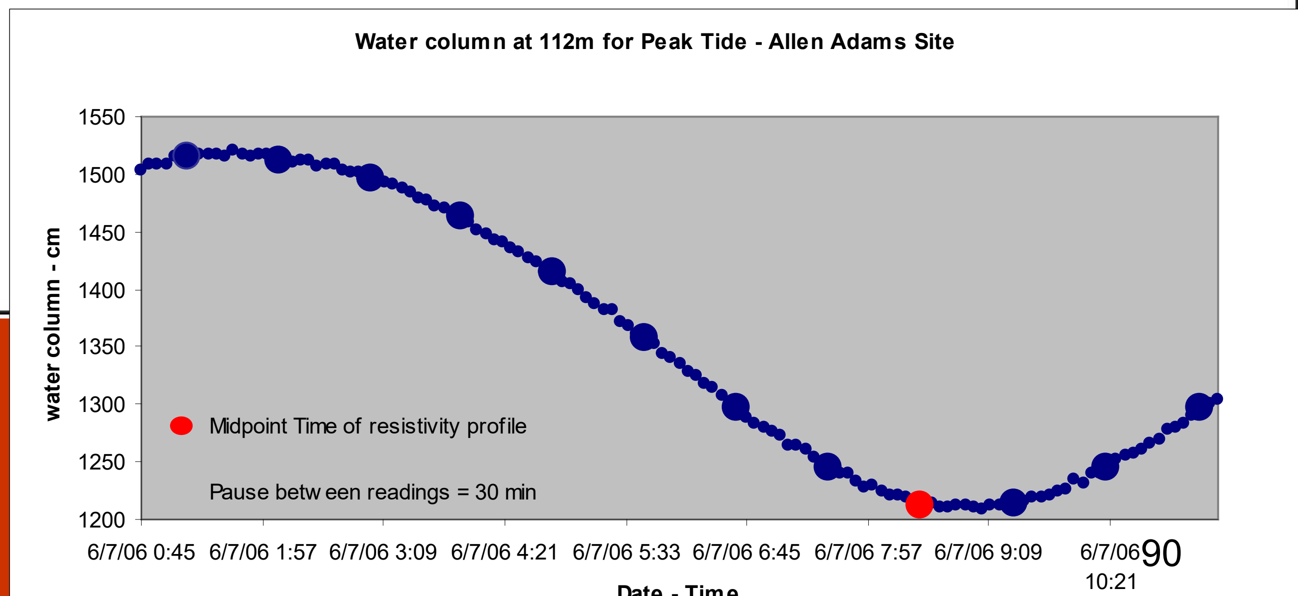






Courtesy of Bill Simonds

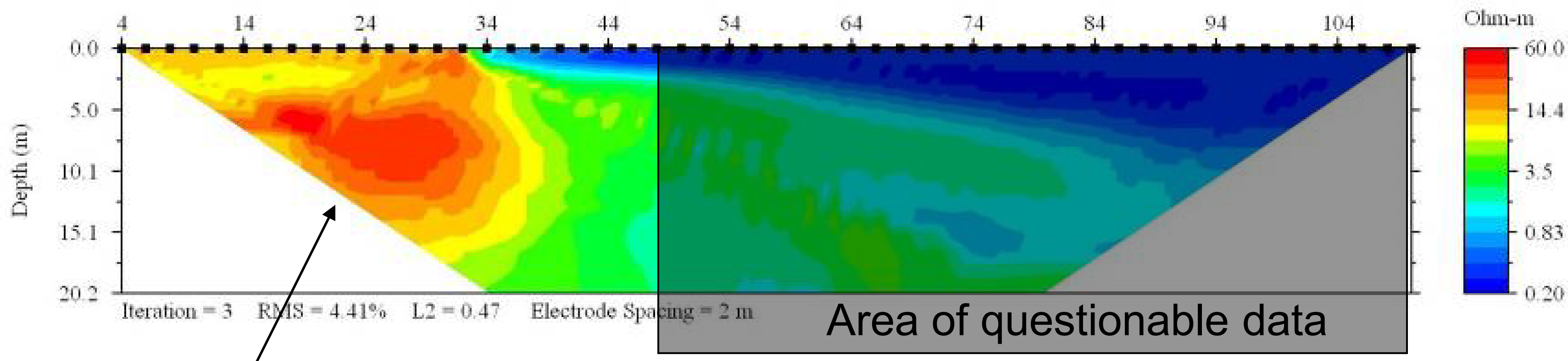
8:29



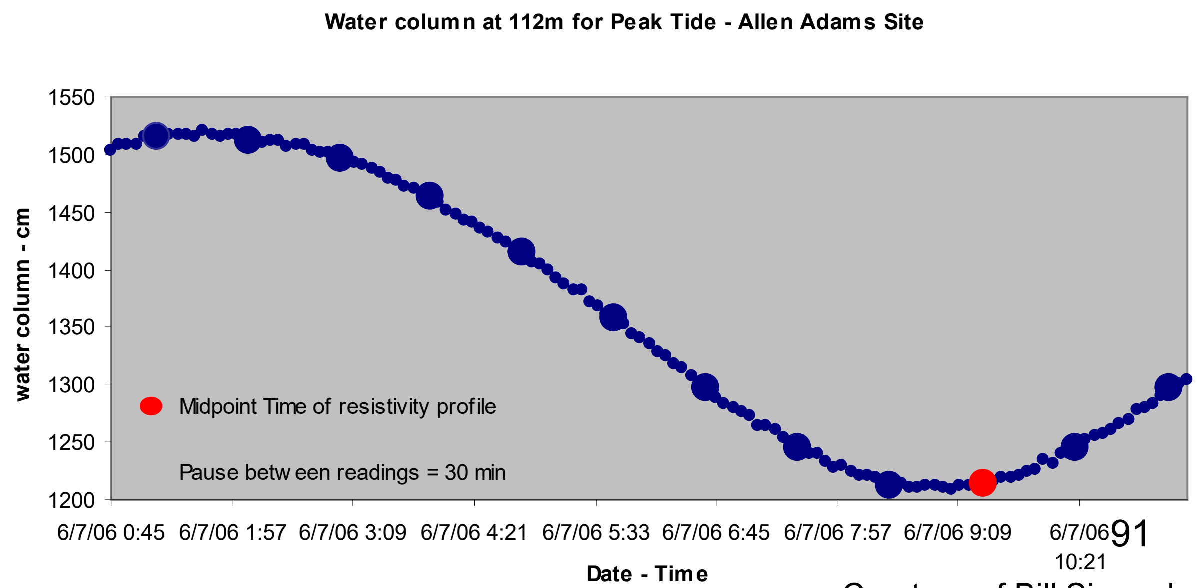




Courtesy of Bill Simonds

9:24

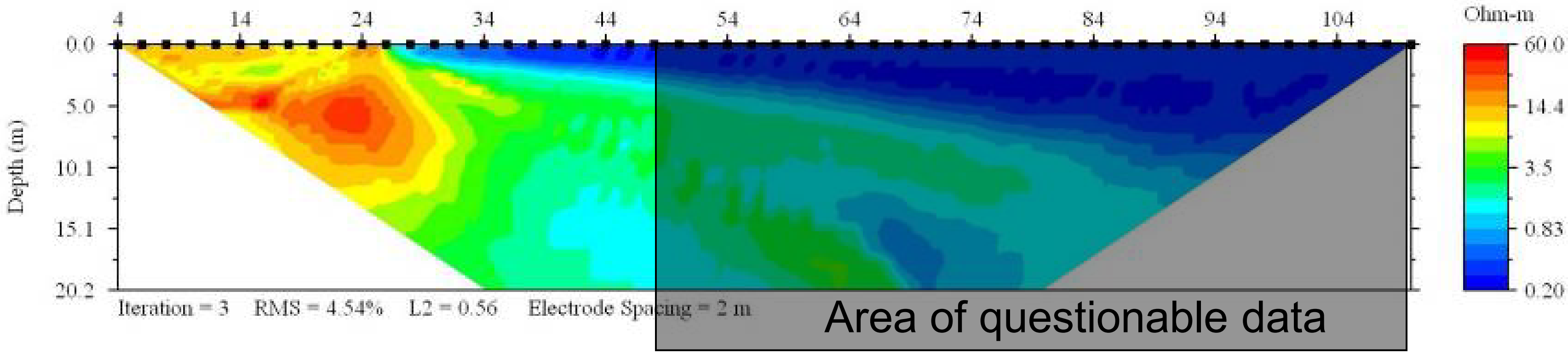



Maximum extent of fresh water

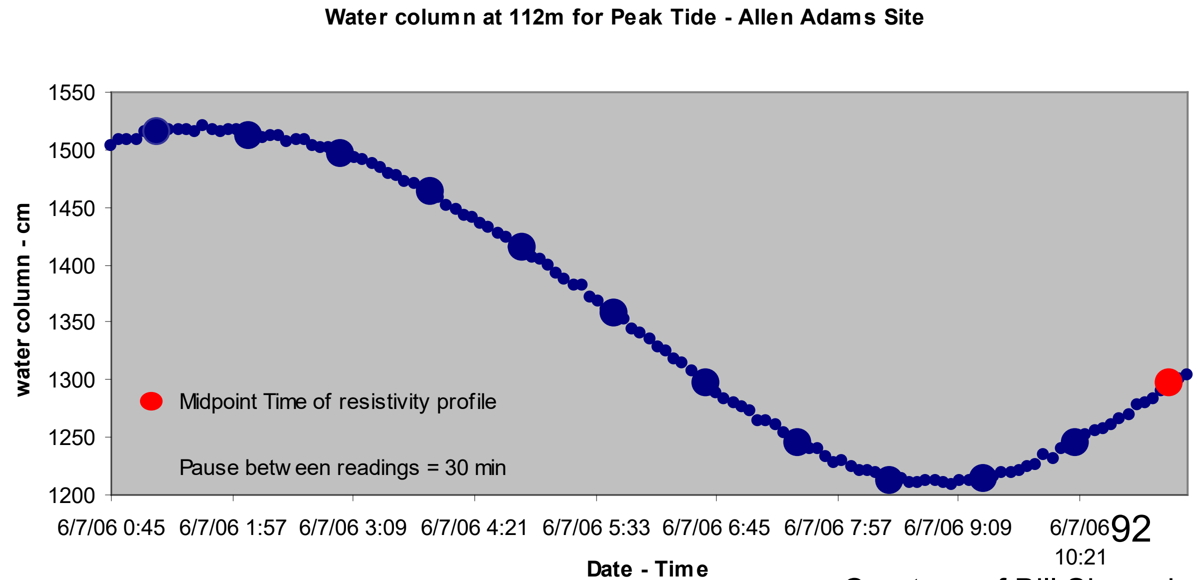



Courtesy of Bill Simonds

# 11:14




Now as the tide begins to rise the sea water is pushing the fresher water in the near-shore sediments back toward the shoreline.






Courtesy of Bill Simonds

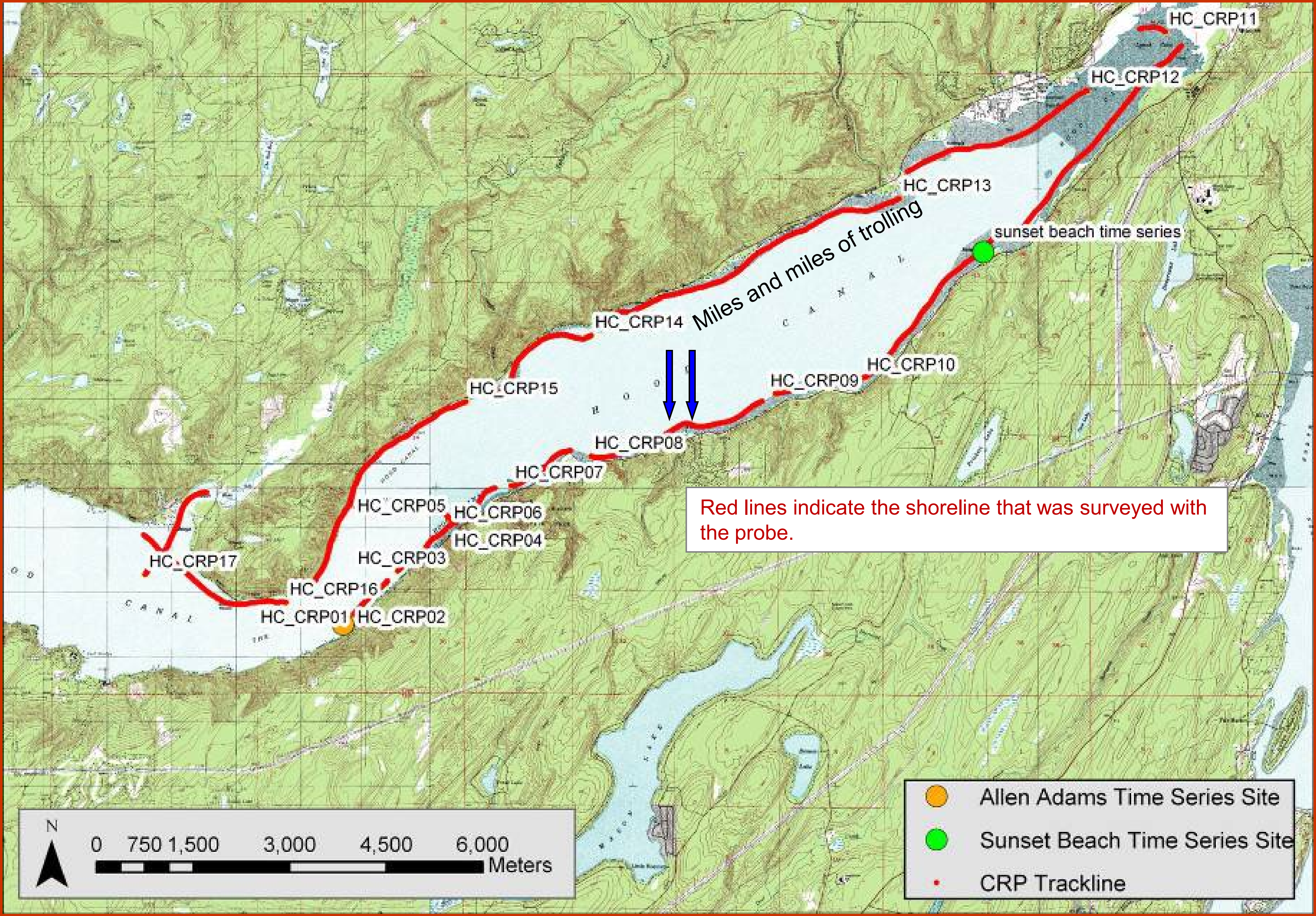
# Collected Streaming Resistivity Profiles and continuous Radon along a transect parallel to the shoreline



Here we are towing the cable on the surface to map areas of fresher water in the sediments beneath the sediment-water interface.

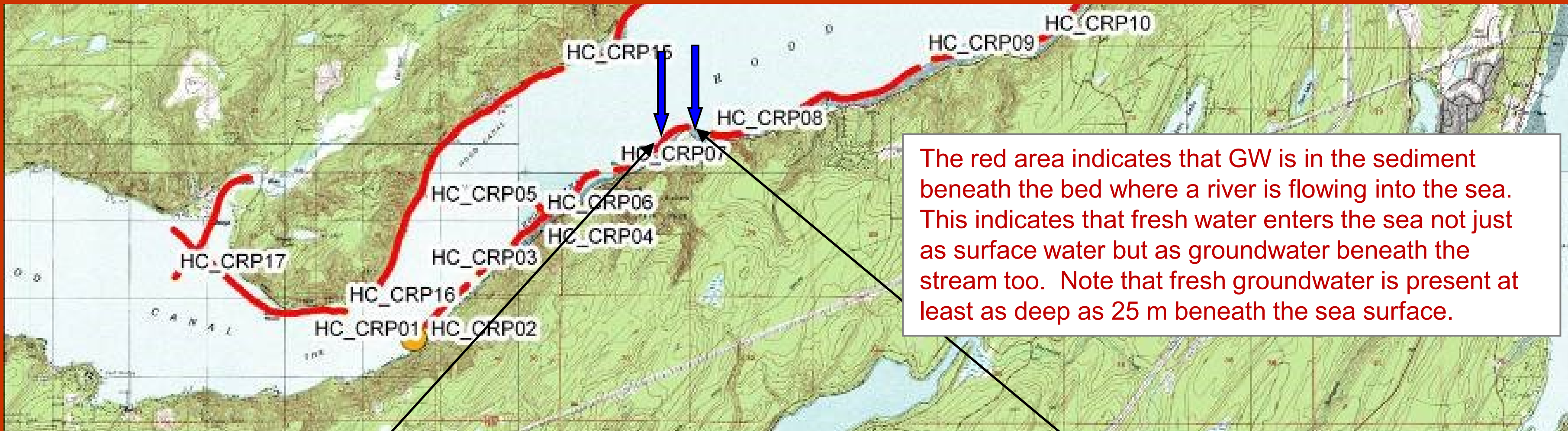


Courtesy of Bill Simonds



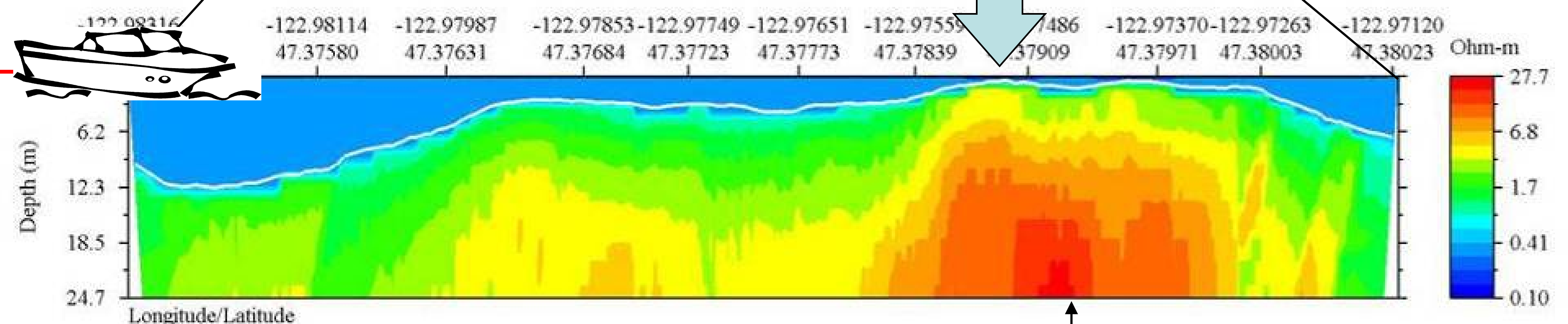



Courtesy of Bill Simonds




The red area indicates that GW is in the sediment beneath the bed where a river is flowing into the sea. This indicates that fresh water enters the sea not just as surface water but as groundwater beneath the stream too. Note that fresh groundwater is present at least as deep as 25 m beneath the sea surface.

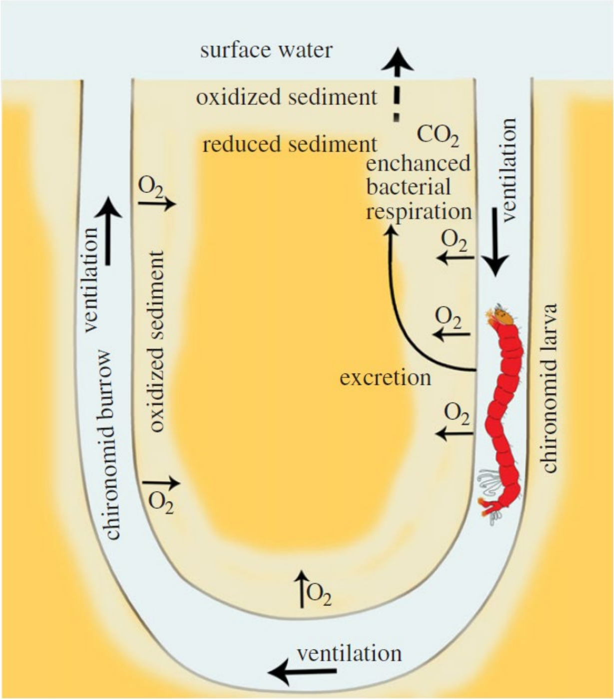
TWANOH CREEK




Resistive fresh-water plume from creek



Simonds et al, 2008, USGS




# Biological indicators

- Presence or absence of some species indicates GW discharge
  - e.g., Marsh marigold
- Bioturbation
  - Benthic invertebrates can manipulate sediment
  - In some cases, burrow tubes and fecal pellets have higher *K* than sediment (e.g., Nogaro et al., 2006, Freshwater Biology)
  - Fish work the sediment, disturb algal growth, contribute to sediment transport (e.g., Statzner, 2003, WRR)
- Bioirrigation
  - Filter feeders can create their own seepage

Chironamids pump one entire lake volume per week!

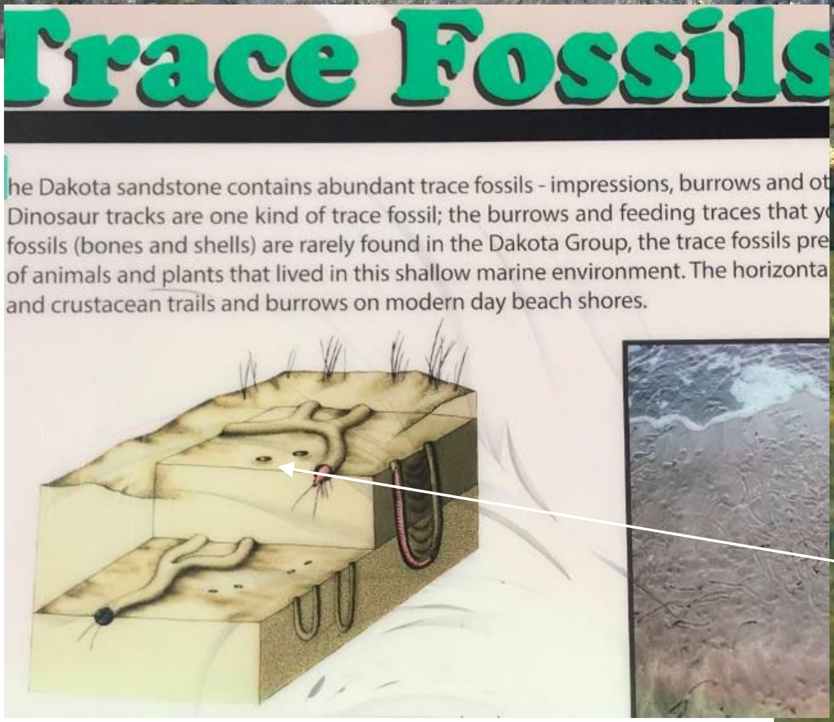
Baranov, Lewandowski & Krause, 2016, *Biology Letters*

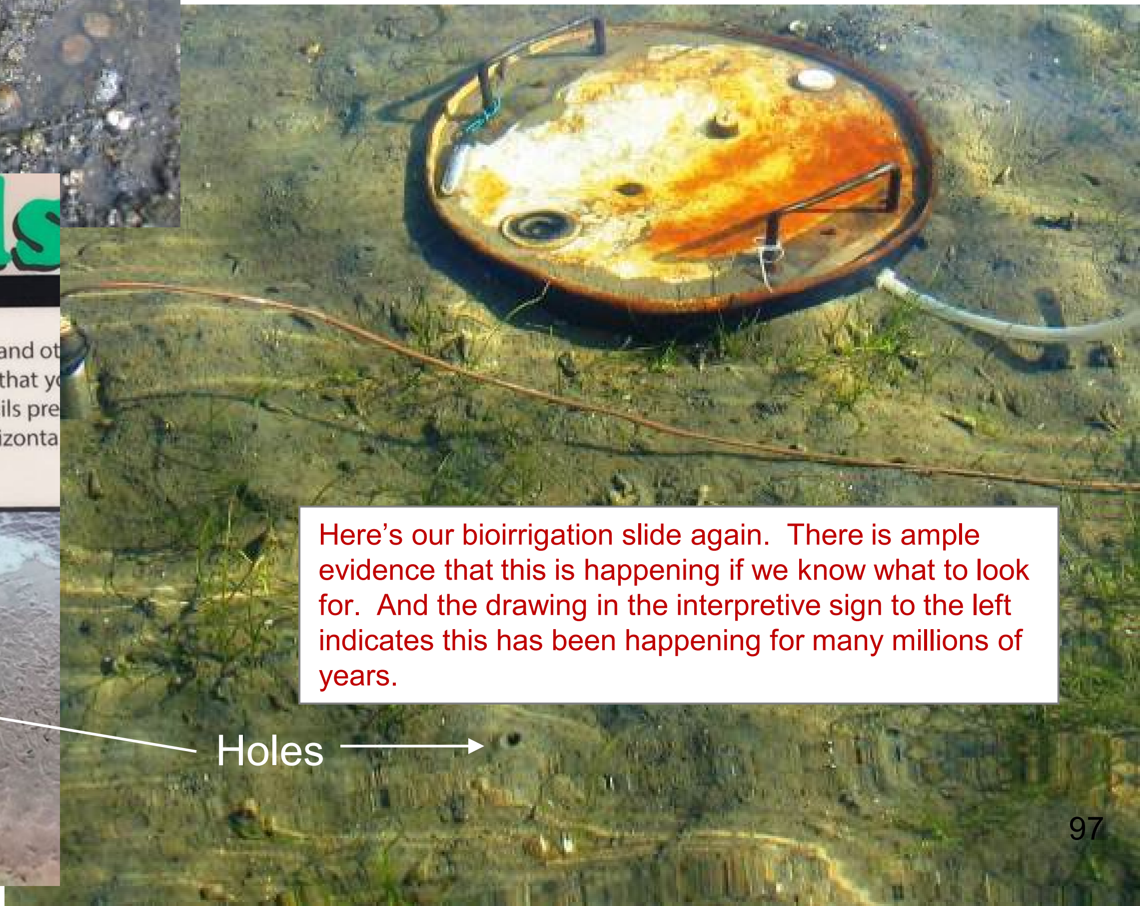
Earlier we talked a bit about chironamids and other filtering animals that live in the bed. We will now talk about these and other biological indicators of areas of focused groundwater discharge.

# Bioirrigation

| Bioirrigating organism | | Linear velocity, cm $day^{-1}$ |
|---|---|---|
| Common name | Species | |
| Ghost shrimp[1] | *Callianassa sp.* | 0.2 |
| Mud shrimp[2] | *Upogebia affinis* (?) | 2.2 |
| Lugworm[3] | *Arenicola marina* | 1.6 |
| Plumed worm[4] | *Diopatra cuprea* | 1.3 |
| | | 5 |

Cable et al., 2006, *L&OM*


Here's our bioirrigation slide again. There is ample evidence that this is happening if we know what to look for. And the drawing in the interpretive sign to the left indicates this has been happening for many millions of years.

## Upper Klamath Lake, OR

10,000 benthic invertebrates per $m^2$

Filtering at least 4 times the rate of diffusion.

1 cm

Kuwabara et al., 2010, USGS OFR

Leech inside seepage bag

## Freshwater bioirrigation

- Rusty crayfish
- Lakes in Minnesota
- ~25 cm/d
- Watch for holes in bed beneath and adjacent to seepage meter

Crayfish that had been inside seepage bag

Not just animals but plants too can be useful indicators of GW discharge to surface water. Once we learned that marsh marigold was a good indicator of GW discharge, we could easily map areas of focused GW discharge by simply mapping the locations of this highly visible plant.


Rosenberry et al., 2000, *Ground Water*

DOI: 10.1002/hyp.14226

INVITED COMMENTARY

HP TODAY WILEY

## Hydraulic conductivity can no longer be considered a fixed property when quantifying flow between groundwater and surface water

# Temporally variable *K*

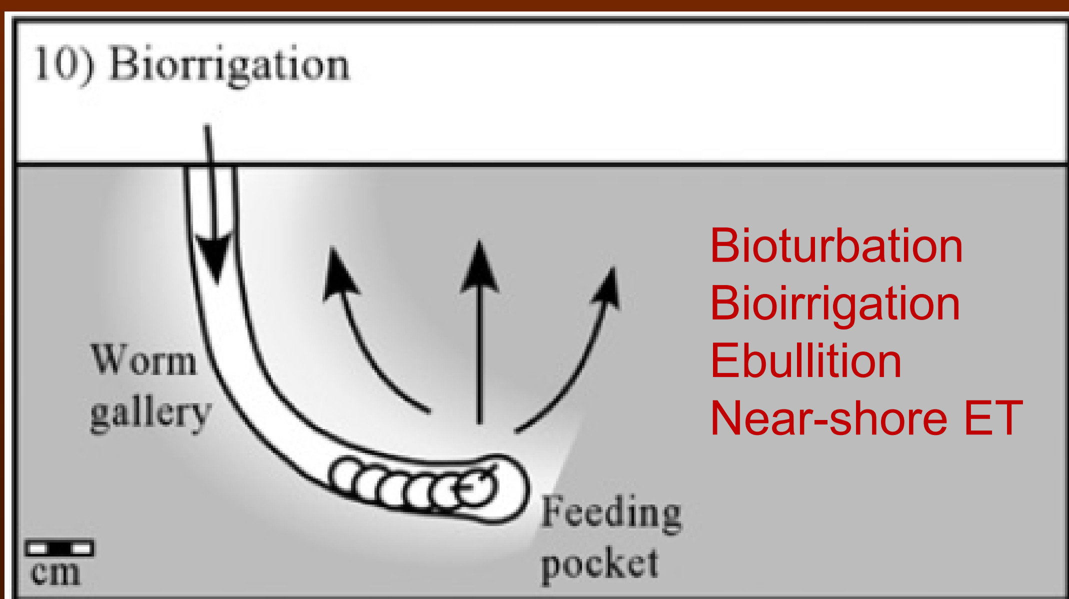
Rosenberry et al., 2021, *Hydrological Processes*

$q = K_K ? i$

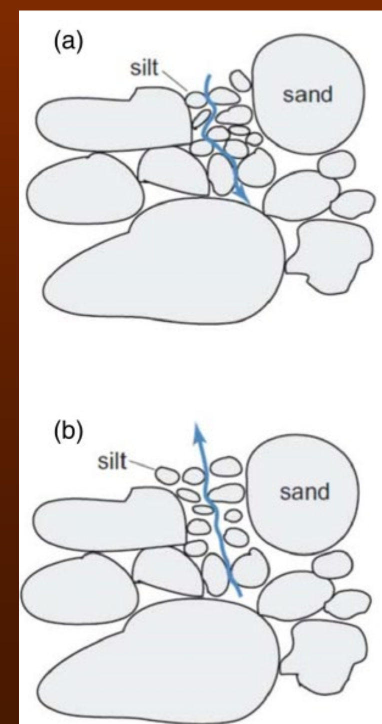
## Why?

1. Episodic deposition and erosion at the bed
2. Flow-directional bias
   Including temperature effects (factor of 2)
3. Biological processes

We've long known that streambeds are constantly evolving, creating changes in *K*. We earlier published the bias related to upward vs. downward seepage, but it was the frequent influence of biological processes that led to our writing this cautionary paper.




Bioturbation
Bioirrigation
Ebullition
Near-shore ET



# Water budget analysis (solve for GW component as the residual of the budget)

$$P + GWI + SWI + OLF - E - GWO - SWO = \Delta LV$$

Can only get net GW →

$$GWI - GWO = \Delta LV + E + SWO - P - SWI - OLF$$

| Terms | |
|---|---|
| *P* | precipitation |
| *GWI, GWO* | GW flow in, out of lake |
| *SWI, SWO* | SW flow in, out of lake |
| *E* | evaporation |
| *ΔLV* | change in lake volume |
| *OLF* | assumed negligible |

| Errors | |
|---|---|
| *P* | 5% |
| *GWI, GWO* | 25% |
| *SWI, SWO* | 5% |
| *E* | 15% |
| *ΔLV* | 10% |

But if we have isotopes or another conservative constituent, we can get *GWI* and *GWO*

$$GWI = \frac{P(C_L - C_P) + E(C_E - C_L)}{C_{GWI} - C_L}$$

$$GWO = \frac{P \cdot C_P - \Delta LV \cdot C_L - GWI \cdot C_{GWI}}{C_L}$$

Now we are switching gears to talk about use of water budgets to calculate groundwater-surface-water exchange. **Masaki also will present an exercise where you will get to do this**. For many lakes, this is the only reasonable option for determining net exchange between GW and SW.

Rosenberry and Hayashi, 2013, Wetland hydrology textbook chapter

# First-order error analysis

$$Error_{RHS} = \sqrt{\overset{\text{SW in}}{5^2} + \overset{\text{SW out}}{5^2} + \overset{\text{Ppt}}{5^2} + \overset{\text{Evap}}{15^2} + \overset{\Delta\text{ stor}}{10^2}} = 20\%$$

**based on percentage errors for specific water-budget components**

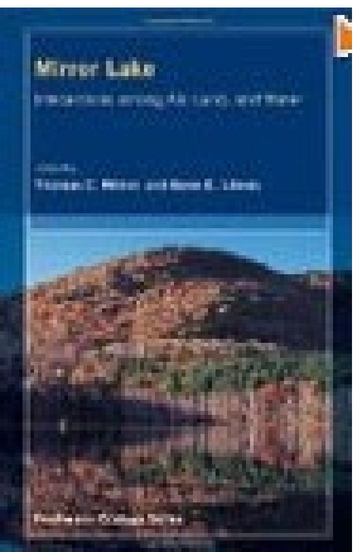
That's better than the 25% error for measuring the GW terms

- This assumes all terms are ~ equal
  - Results would be better for SW-dominated system
- GW estimates are based on lots of wells and repeat measurements – may actually be closer to 50% if we don't know the geology

In many cases, it is simply not cost effective to measure *GWI* or *GWO* using networks of wells, especially if we can do better by simply measuring the other components

- Still need a few wells for determining the conservative constituent in the ground water upgradient and downgradient of the surface-water body, though.

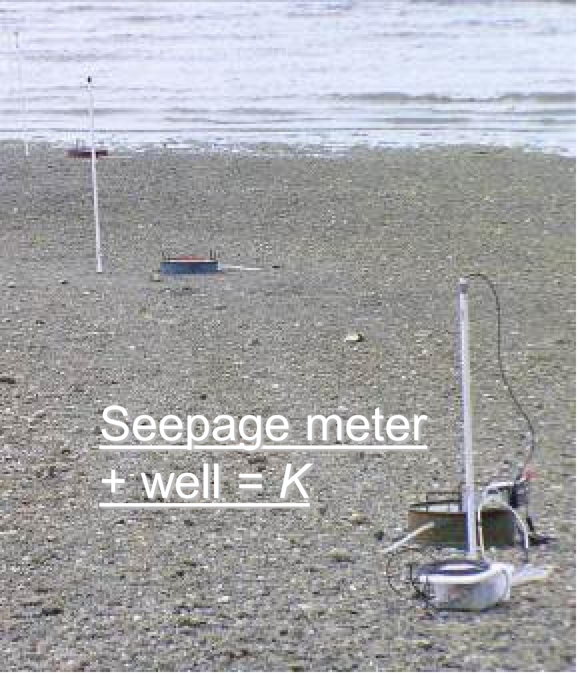
| This is a great method that works well if we can keep the errors of our water budget terms relatively small. |
|---|

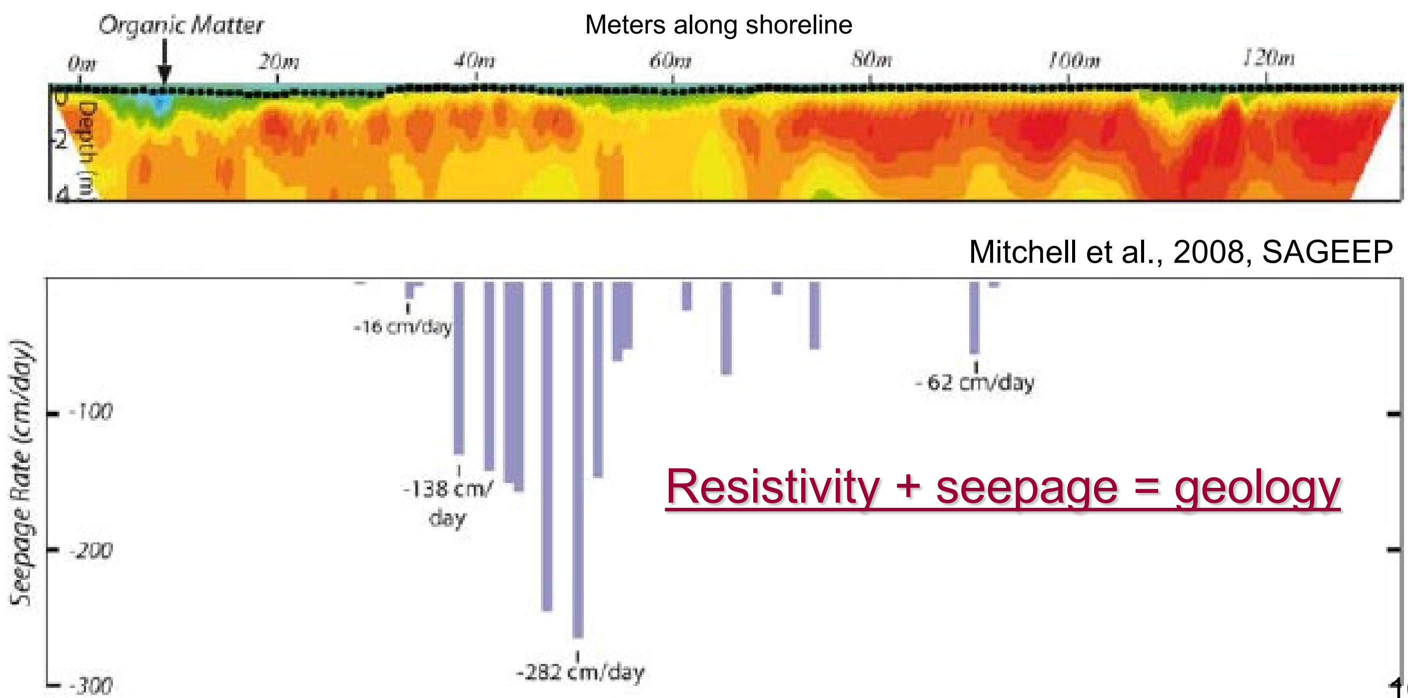



Mirror Lake book on water and chemical budgets, 2009, Univ. of California Press

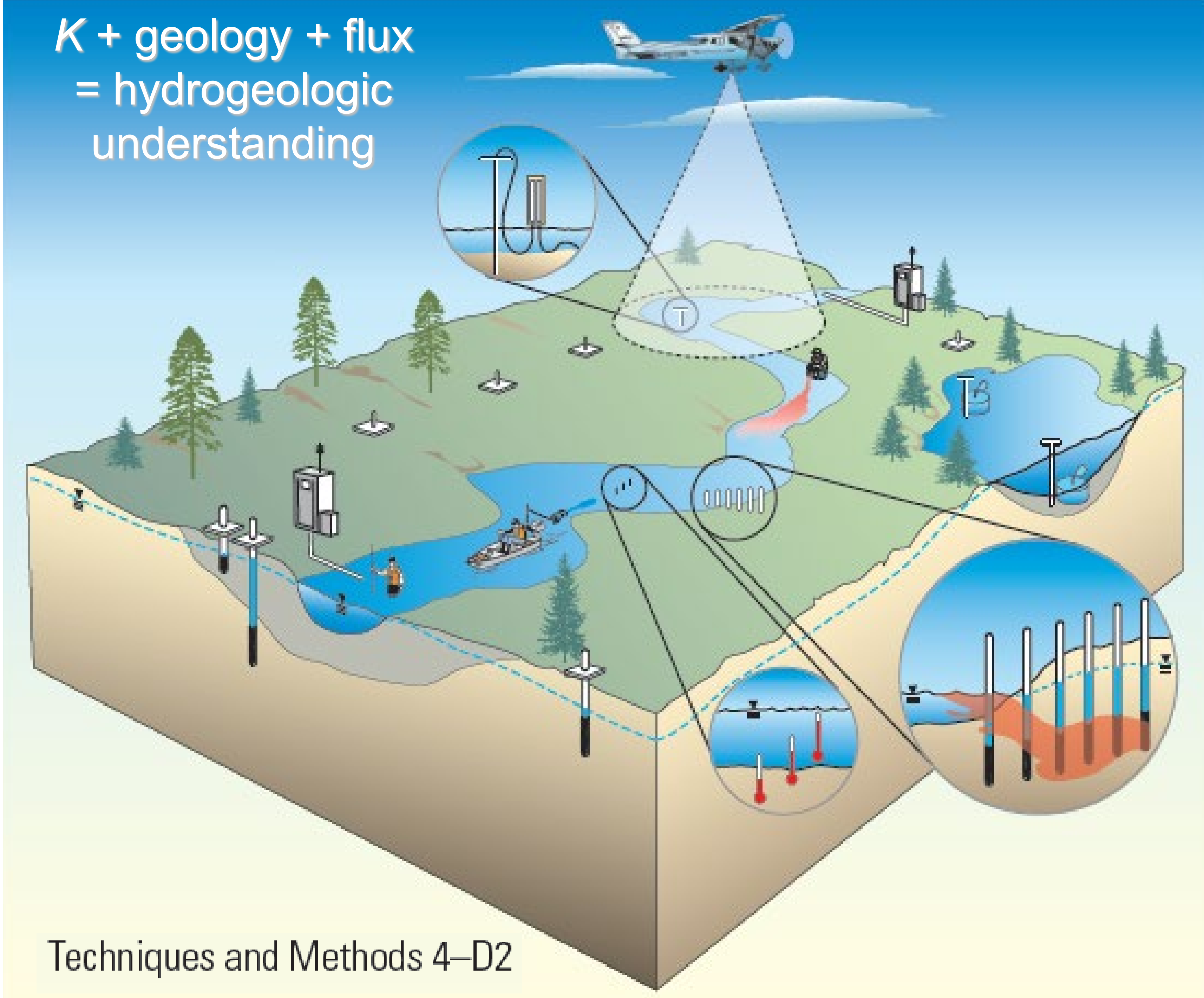
Take-home message. . .

## Use multiple methods



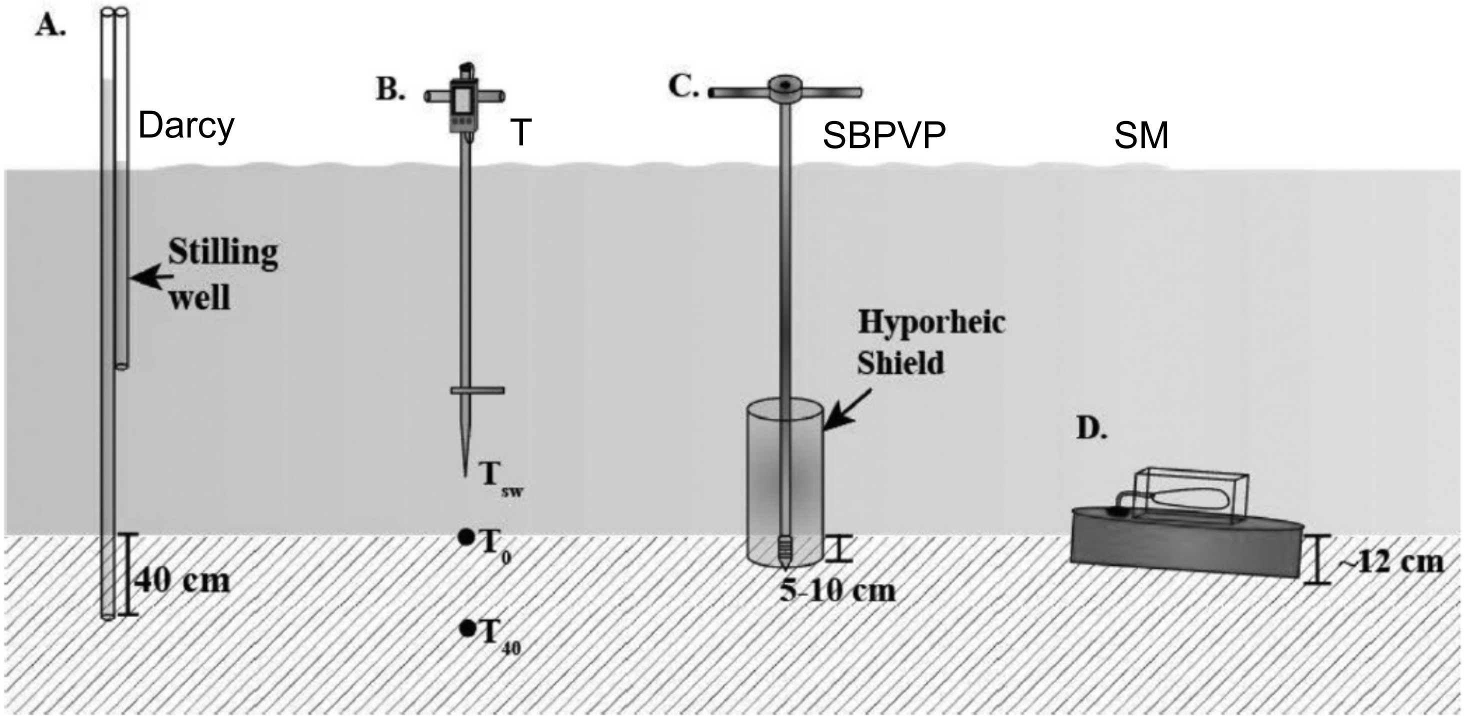



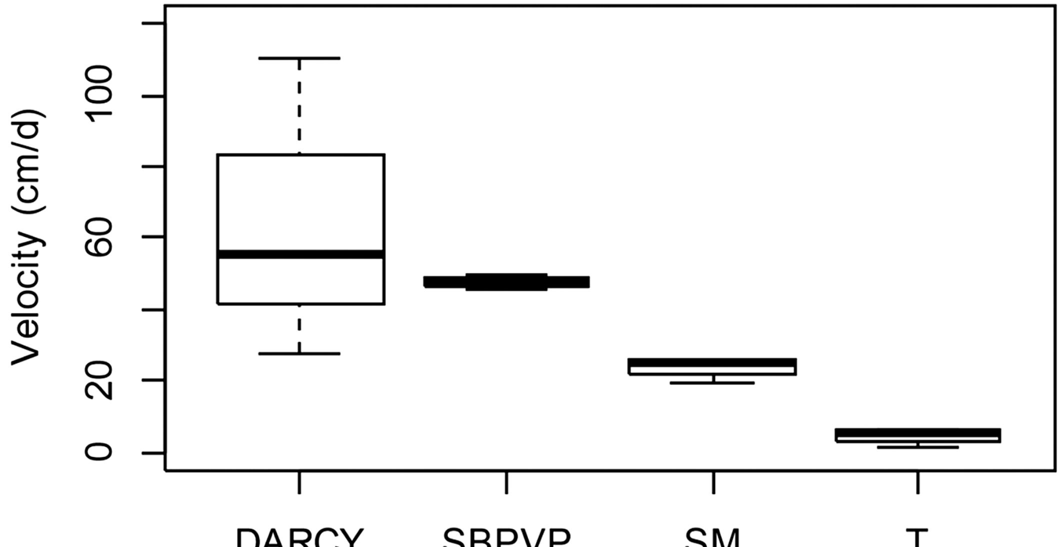
K + geology + flux
= hydrogeologic
understanding



Techniques and Methods 4–D2

# And this word is getting out. Here's a recent example.




Here's a recent example of use of multiple methods in a streambed in Denmark. The **Darcy** method measures the vertical hydraulic gradient and requires an estimation of K. The **T**emperature method is based on the Schmidt et al., 2007, method. The **S**tream**b**ed **P**oint **V**elocity **P**robe is a tracer-based method. The **S**eepage **M**eter makes a direct measurement of the flux across the sediment-water interface.

Cremeans and Devlin, 2020, *GW Mon. & Remediation*

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