Università di Trieste
Corso di Laurea in Geologia

Anno accademico 2015 - 2016

Geologia Marina

Parte IV

Modulo 4.1 Cause e modalità del movimento di fluidi nei sedimenti

Docente
A. Camerlenghi

In an ideal sedimentary column where the pore pressure field is hydrostatic, the temperature distribution is controlled by conduction of heat (no advection) and there are no chemical gradients (e.g. salinity, osmosis), the pore fluids do not move. Better: they do move upwards as the porosity decreases with time, but they remain in the sedimentary column.

Net pore fluid movement from the subsurface to through the seafloor is triggered by the establishment of pore **pressure**, **thermal**, **chemical** gradients that depend on the **sedimentary**, **tectonic** and **diagenetic** histories of the sedimentary basin

During the process of consolidation, pore fluids are expelled, initially into the pores of the surrounding sediments

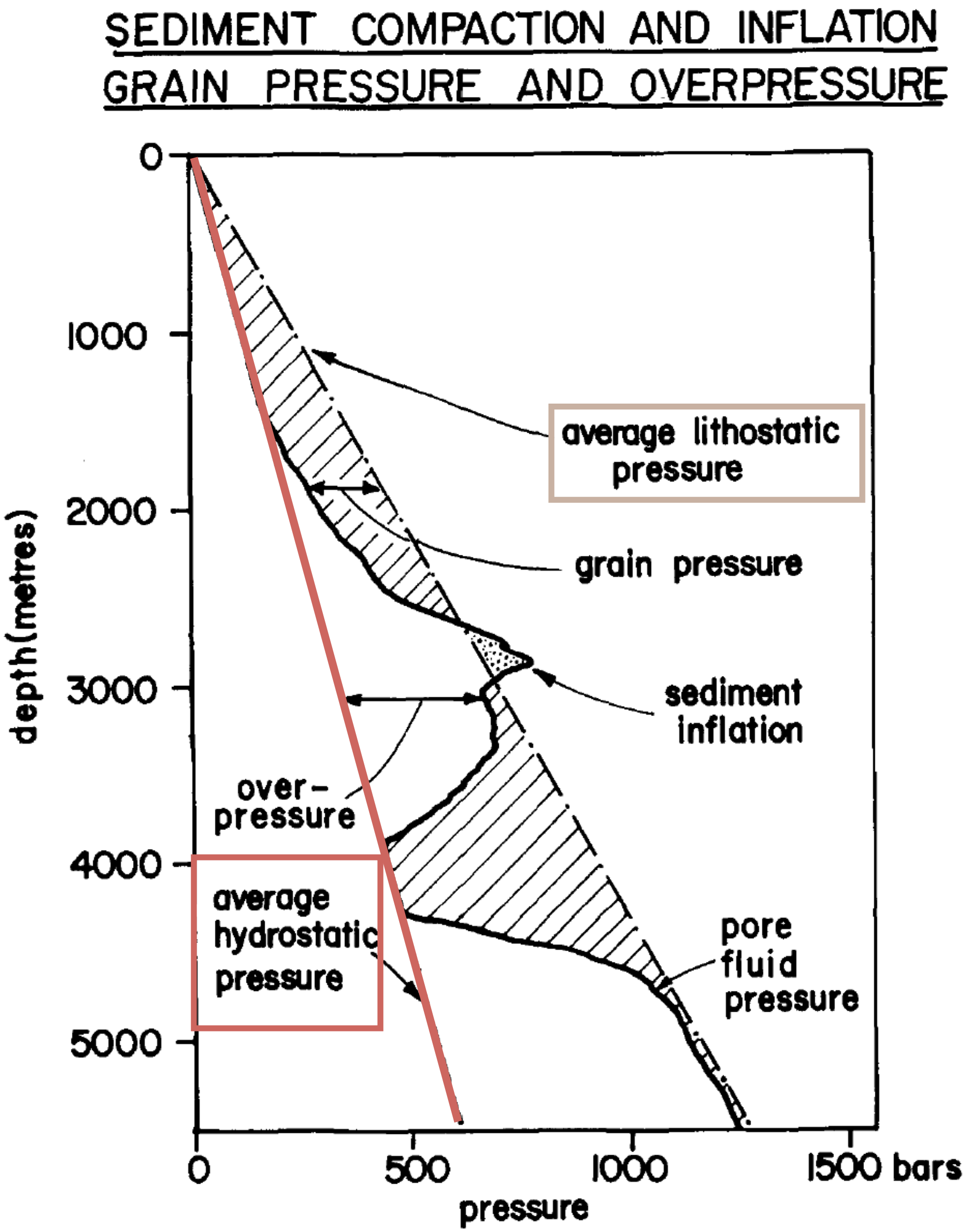
In homogeneous sedimentary columns fluids move up-wards, towards strata with higher porosity >>>> higher permeability.

In inhomogeneous sedimentary columns sands and oozes experience little compaction (they retain more fluids during compaction).
Clays undergo a much larger porosity reduction during compaction (e.g. From 80% to 40 % porosity reduction).
Fluids may move from clay-rich strata to sandy strata by moving laterally, upwards, and downwards.
Alternatively fluids may move using fractures and faults.

Compaction Disequilibriu

Rapid loading by a huge thickness of the same clay/silt sediment may tip the dewatering balance temporarily in favour of overpressure. In fact the dewatering process is rarely perfectly "normal".

This lack of dewatering conspires to cause the matrix stress between the grains to become "locked" as burial continues, and causes the pore fluids to be responsible for carrying the remaining overburden.



Hydrostatic Gradient (not counting seawater Compressibility) 9.81 ~ 10 kPa m^{-1}

Jeans, 1986

Compaction Disequilibrium

The process will continue until the fluid pressure finds relief by rupturing the seal.

This rupture can occur at pressures below the overburden if the rock is brittle or even as much as 40% above the overburden if the rocks have enough tensile strength.

Since **compaction disequilibrium** is common in younger clays, a frequent result of this effect is a suite of **mud diapirs**, **mud lumps**, and **sand volcanoes**.

Pore pressure gradients

Basic concepts:

A soil (marine sediment in our case) is a compressible skeleton of particles.

- *particles are not compressible*
- *the interstitial fluid (water, gas, or both) is compressible*
- *the skeleton is compressible*

Volume changes (deformations) do not depend on the total stress applied. They depend on the DIFFERENCE between TOTAL STRESS and PORE FLUID PRESSURE.

$$\Delta V/V = -C_c\,(\Delta\sigma - \Delta u)$$

ΔV = volume change (being V the original volume)
$\Delta\sigma$ = change in total stress
Δu = change in pore fluid pressure
C_c = compressibility of the soil skeleton

A volumetric change can be obtained by an increase in $\Delta\sigma$ as well as by a decrease in Δu

Sediment consolidation state

Consolidation is the volumetric change induced by a change of total stress (sedimentary burial, tectonic burial).

When a sediment is buried, it undergoes a process of natural consolidation due to the progressive increase of the overburden (total stress).

NORMALLY CONSOLIDATED
The overburden is the maximum stress ever experienced.

OVER-CONSOLIDATED
The overburden is the less than the maximum stress ever experienced.

UNDER-CONSOLIDATED
The overburden is more than the maximum stress ever experienced.

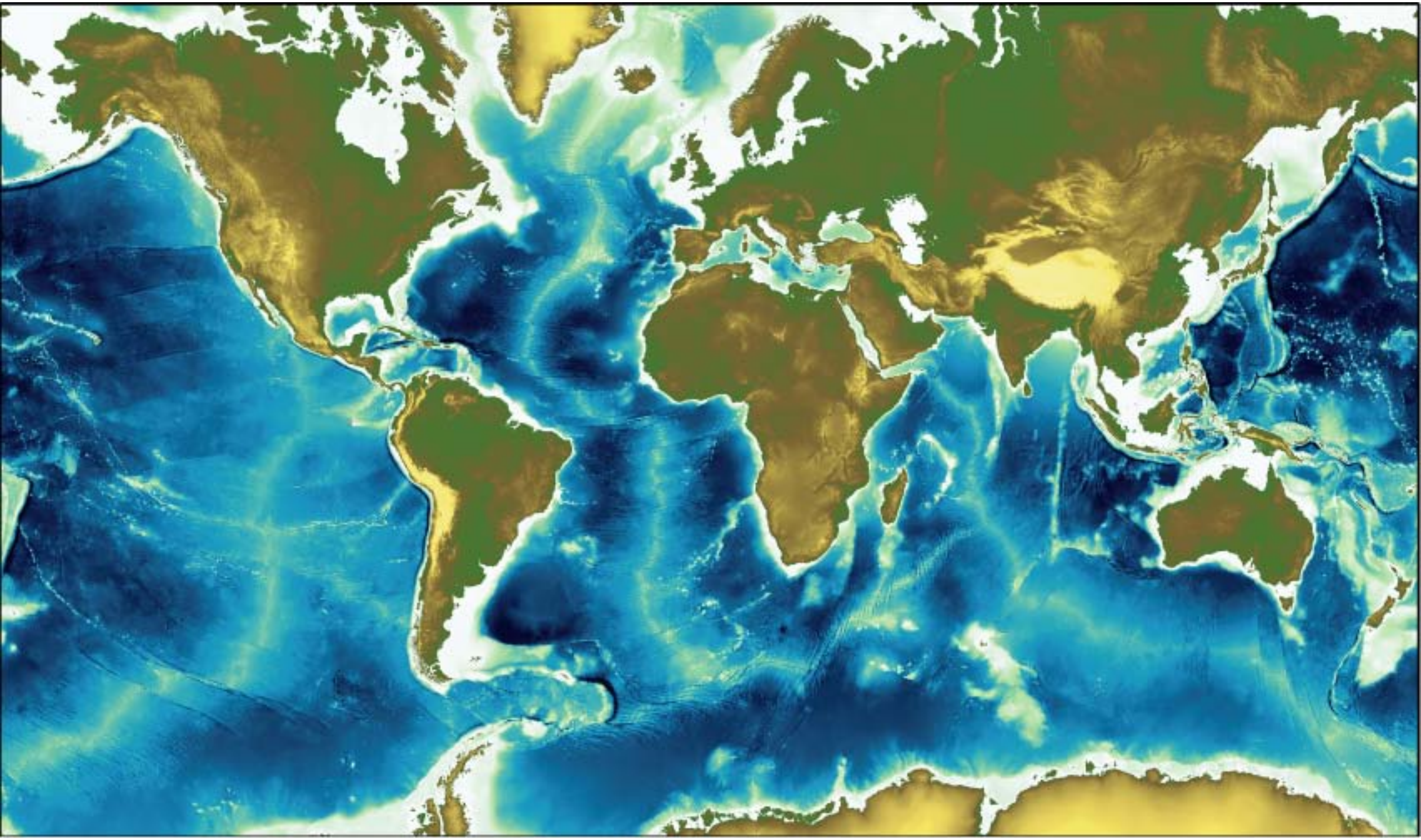
Factors that determine excess pore water pressure and influence the consolidation state

The most common factors are:

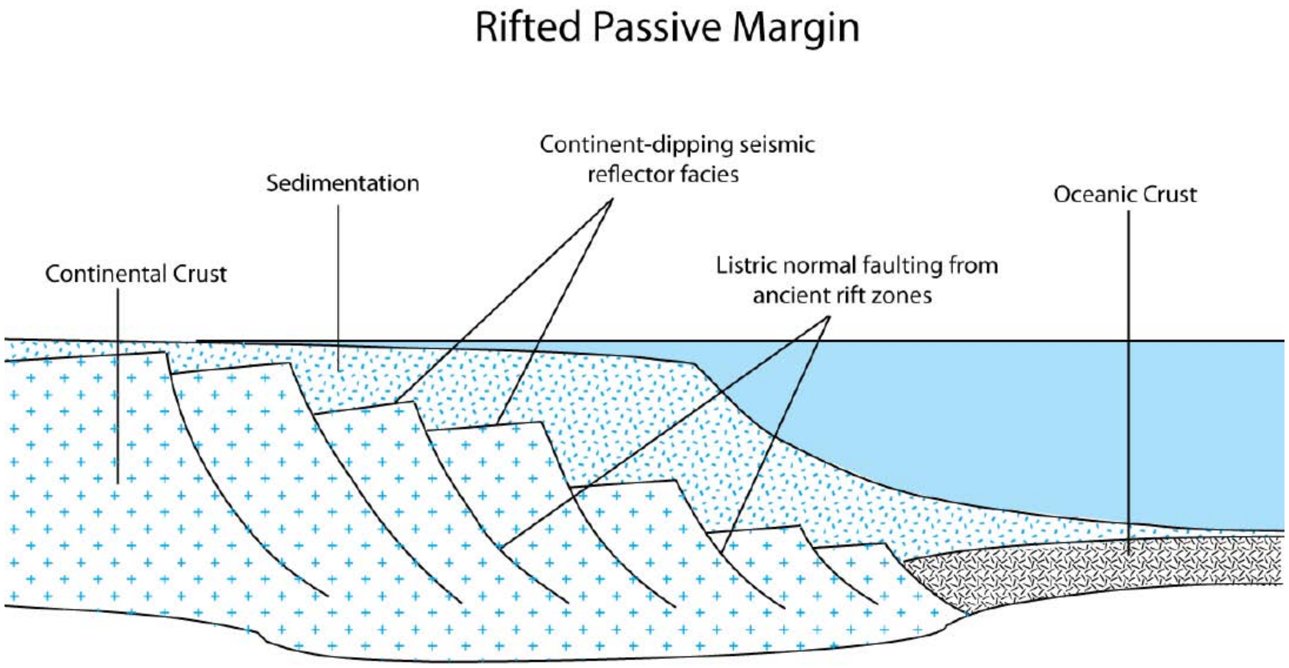
- **Stress-related factors**
    - Rapid sedimentation rate
    - Tectonic stress
- **Fluid volume increase mechanisms**
    - Temperature increase
    - Mineral transformation
    - Hydrocarbon Generation
- **Fluid movement mechanisms**
    - Osmosis
    - Hydraulic head
    - Hydrocarbon buoyancy

(**Swarbrick and Osborne, 1998**. Bryant et al., 1974; Sangrey, 1977; Arthur et al., 1980; Demaison & Moore, 1980; Bryant et al.,1981).

Rapid sedimentation rates

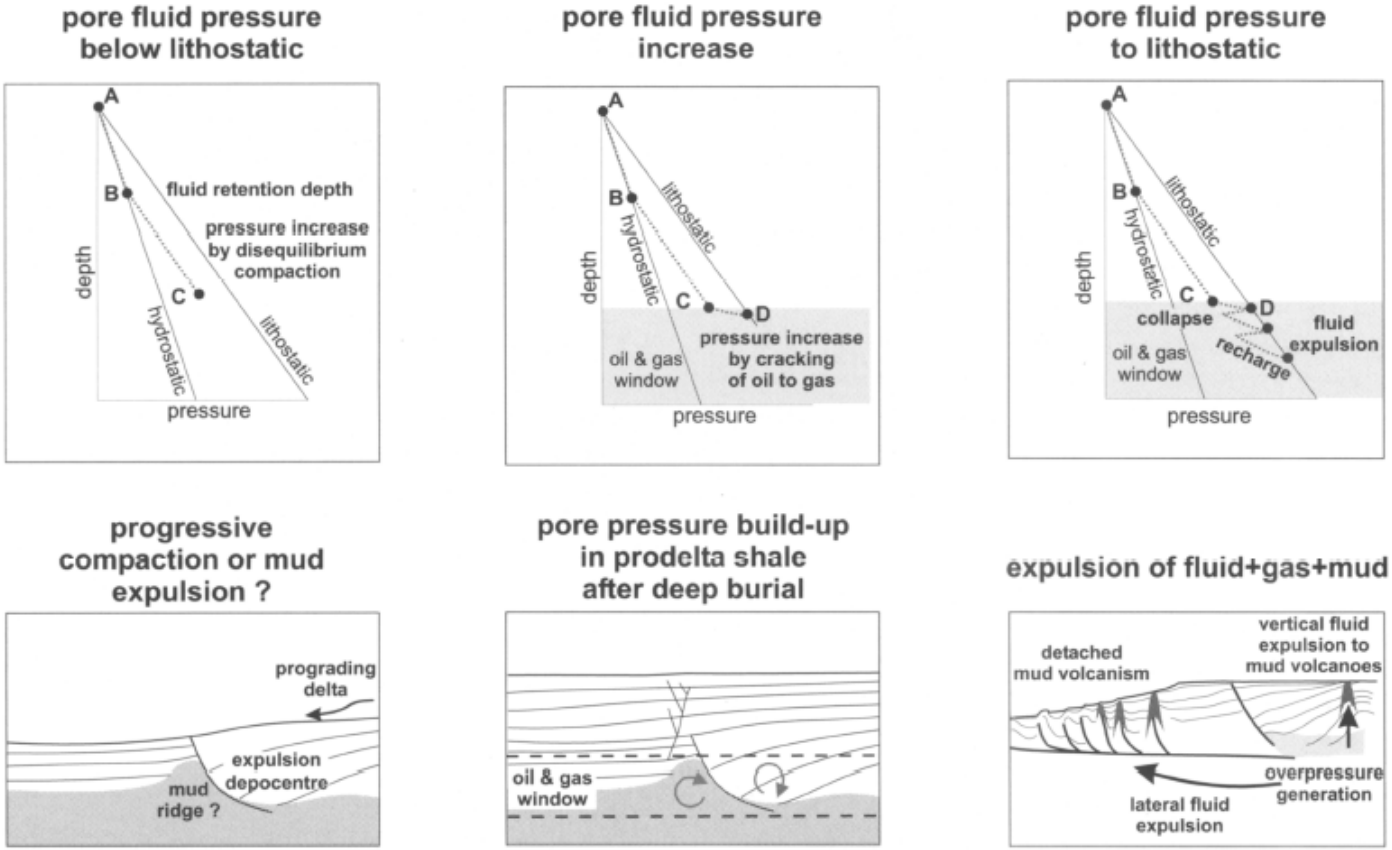


Rapid sedimentation rates



Rapid sedimentation can be produced by:
- **River-dominated** sedimentary systems (low and medium latitudes)
- **Glacial-dominated** sedimentary systems (high latitudes)

Rapid sedimentation rates in river-dominated environments

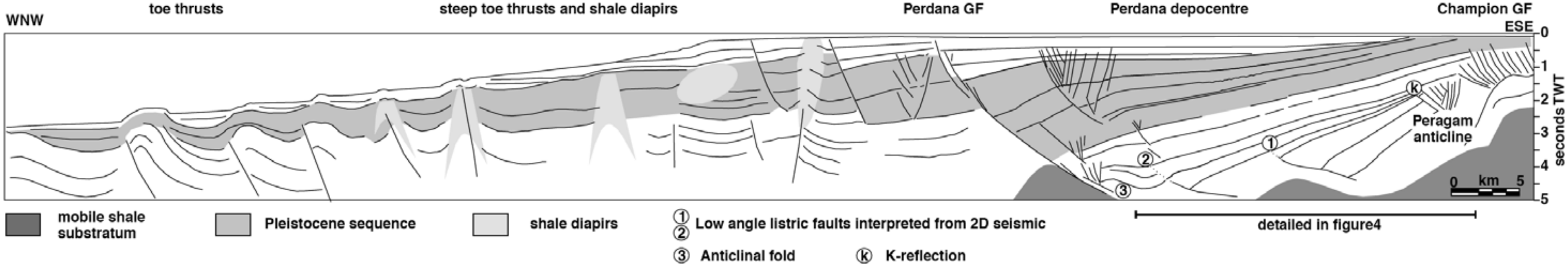


Overview of the styles of shale mobility in relation in relation with pore fluid overpressure; early syndepositional structures can be distinguished from later post-depositional injections of mobilized sediment that occur after large overpressure increase.

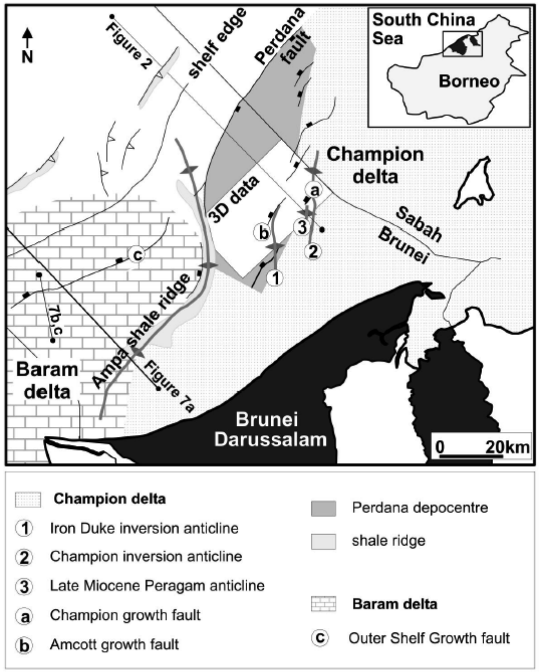
Van Rensbergen and Morley, 2003

Rapid sedimentation rates in river-dominated environments

Shale tectonics, Offshore Brunei



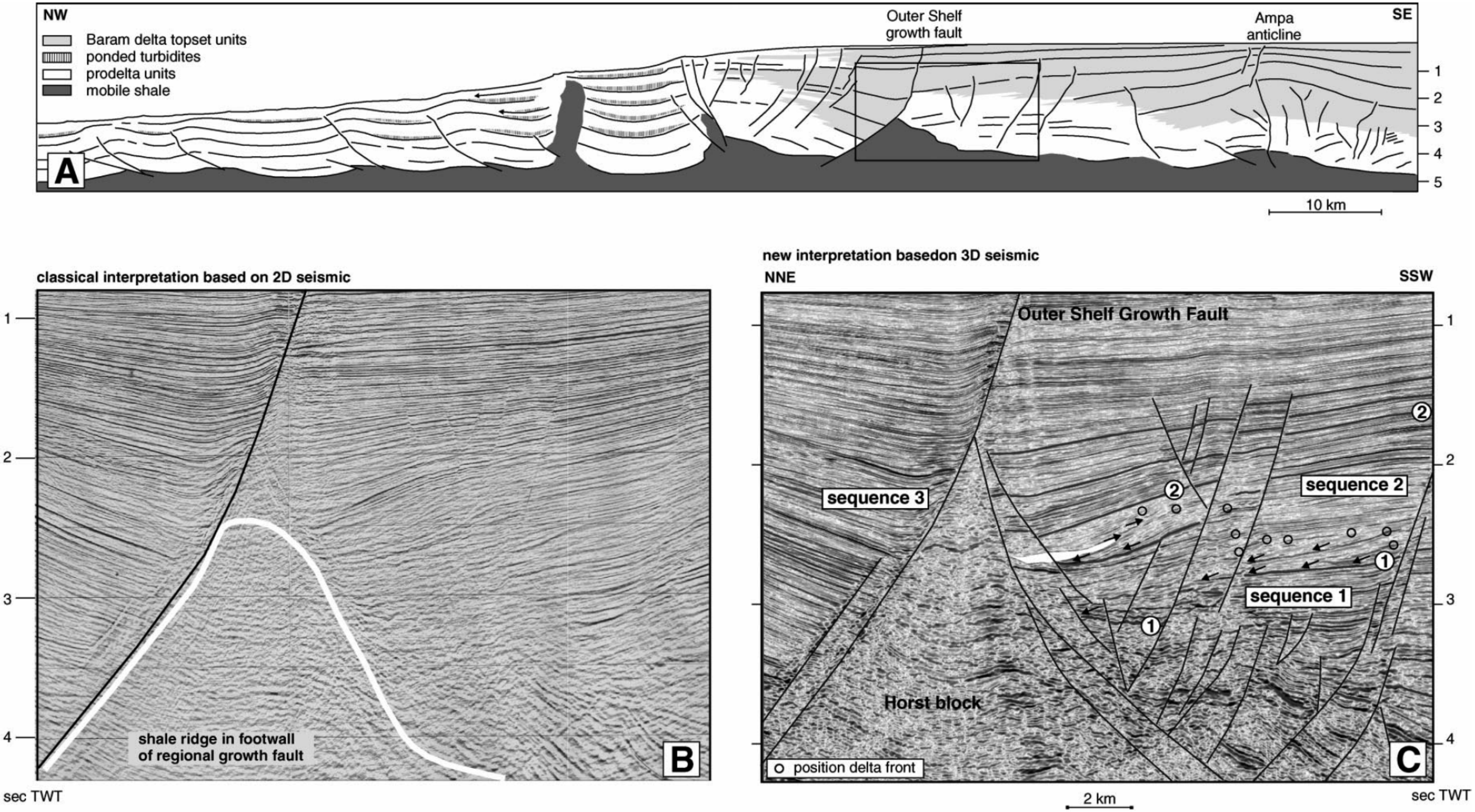
Schematic cross-section of the Champion delta based on 2D regional seismic line.



Van Rensbergen and Morley, 2000

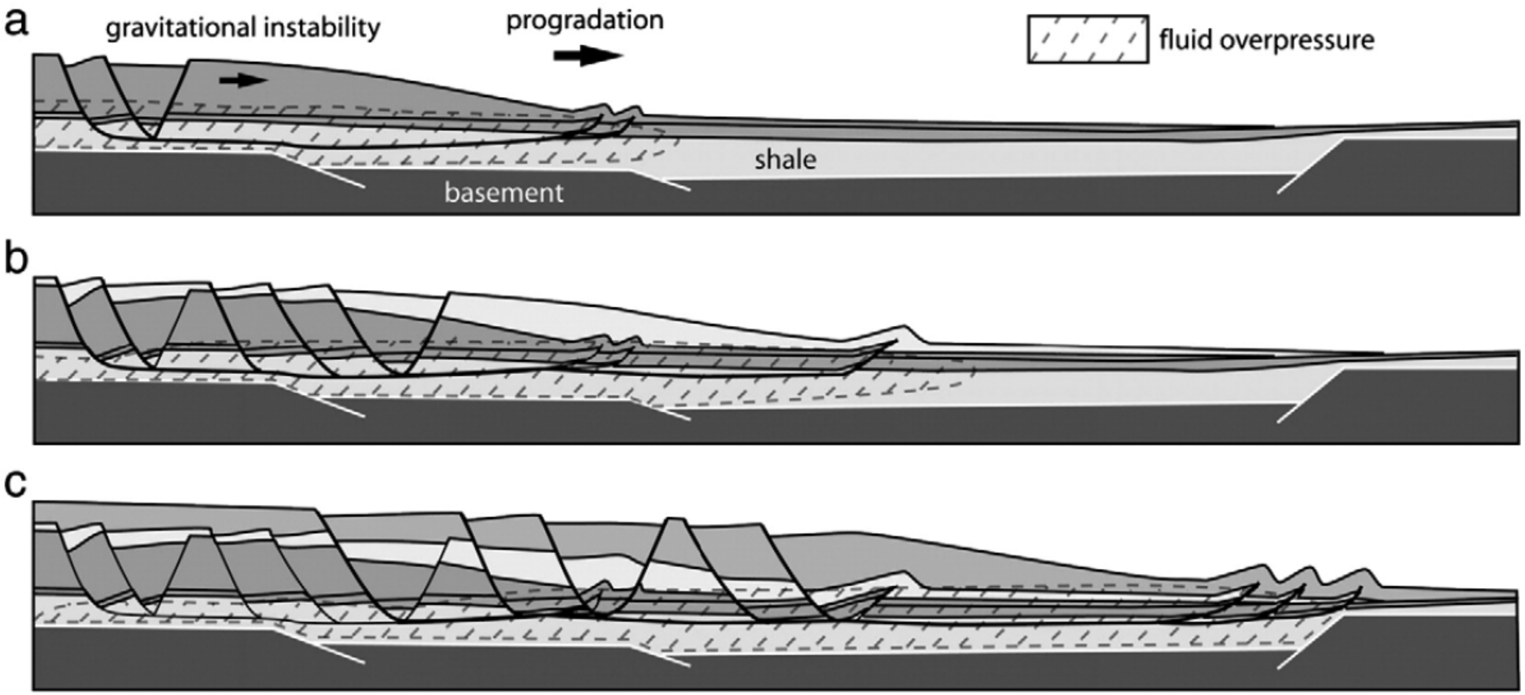
Rapid sedimentation rates in river-dominated environments

Shale tectonics, Offshore Brunei



Van Rensbergen and Morley, 2000

Rapid sedimentation rates in river-dominated environments

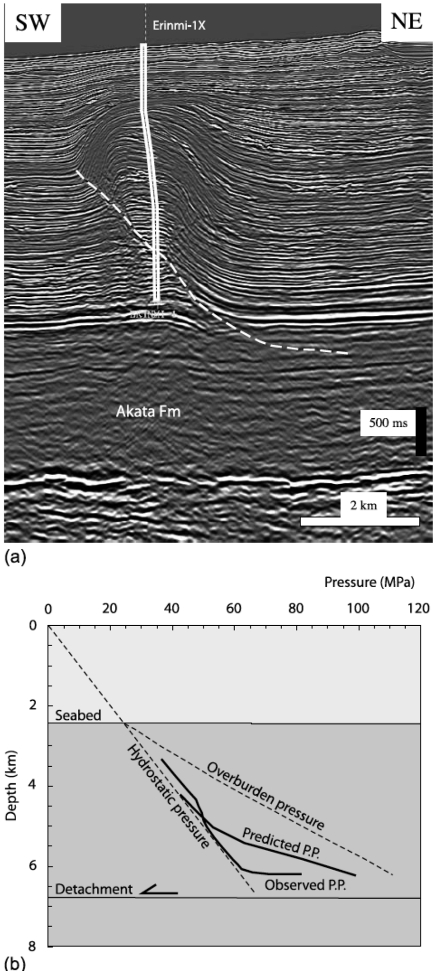


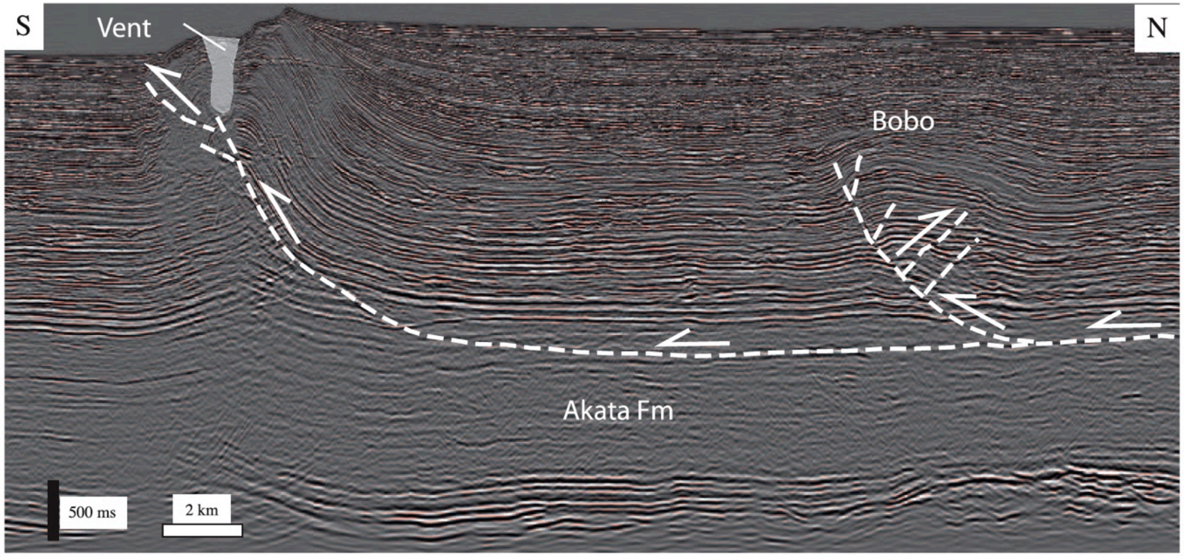
Evolutionary model of fluid overpressure and gravitational deformation in a prograding delta. Whatever the generating mechanism may be (compaction or hydrocarbon generation), the front of the overpressured domain may advance basinward as sediments prograde seaward.

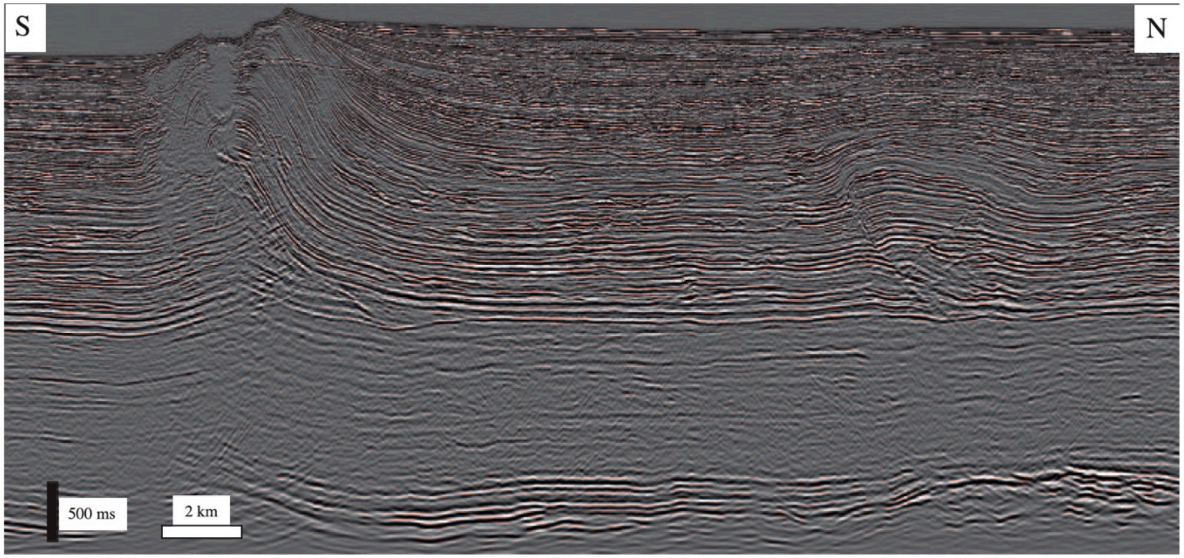
R. Mourgues a,*, E. Lecomte b, B. Vendeville b, S. Raillard, 2009. An experimental investigation of gravity-driven shale tectonics in progradational delta, Tectoophysics, 474, 643–656.

Rapid sedimentation rates in river-dominated environments

Shale tectonics in the Niger Delta





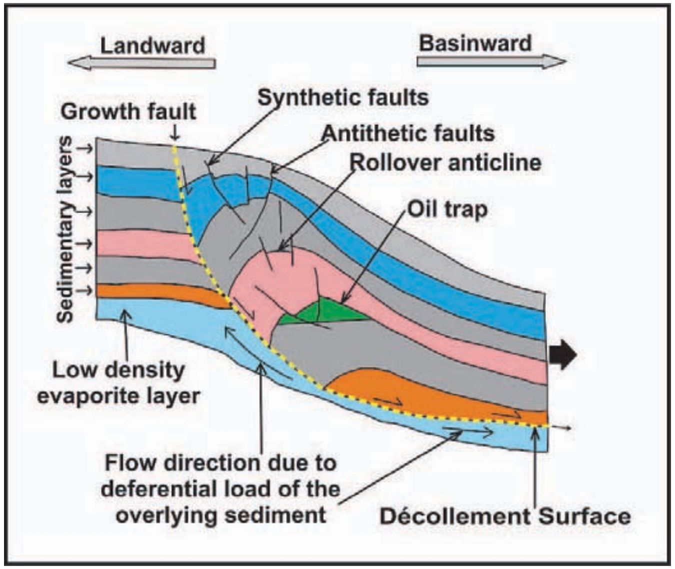


Cobbold&al, 2009. Petroleum Geosciences, 15, 3-15.

Compaction Disequilibrium

Growth Faults

Growth faults are syndepositional or syn-sedimentary extensional faults that initiate and evolve at the margins of continental plates.

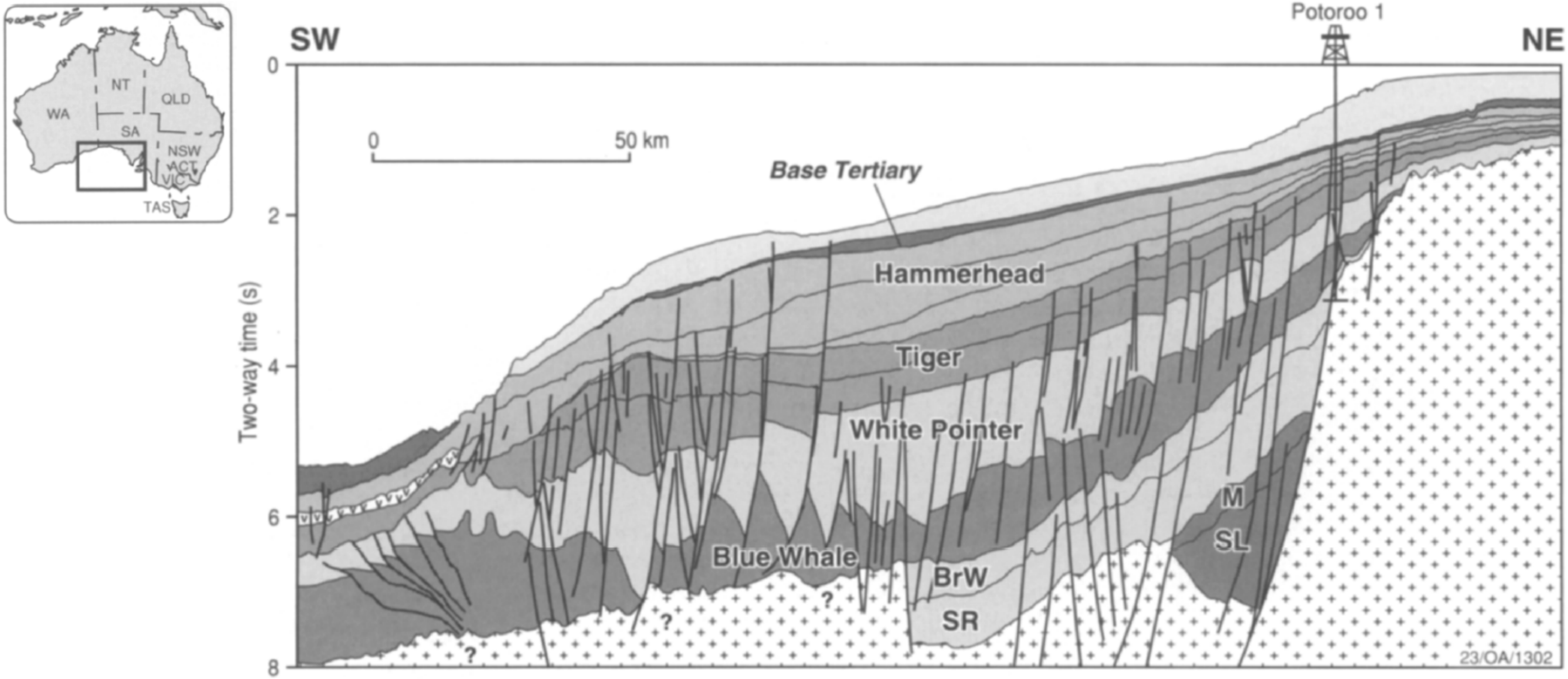


The pressures can sometimes be relieved by systems of **sub-vertical faults** or by **growth fault systems**.

The high pressures in these shale masses are a major contributing factor in the formation of massive "**growth faults**" that cut across the delta, trapping the rollover anticlines (which often form traps for oil and gas in the hanging- wall).

The faults may also trap oil on the foot-wall side where the movement has brought sands against shales to seal them.

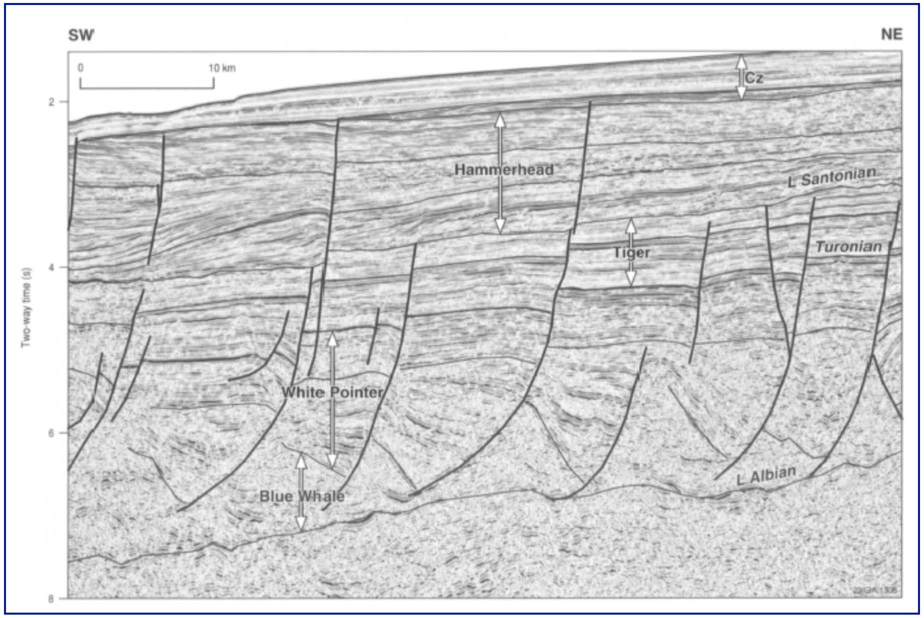
Rapid sedimentation rates in river-dominated environments

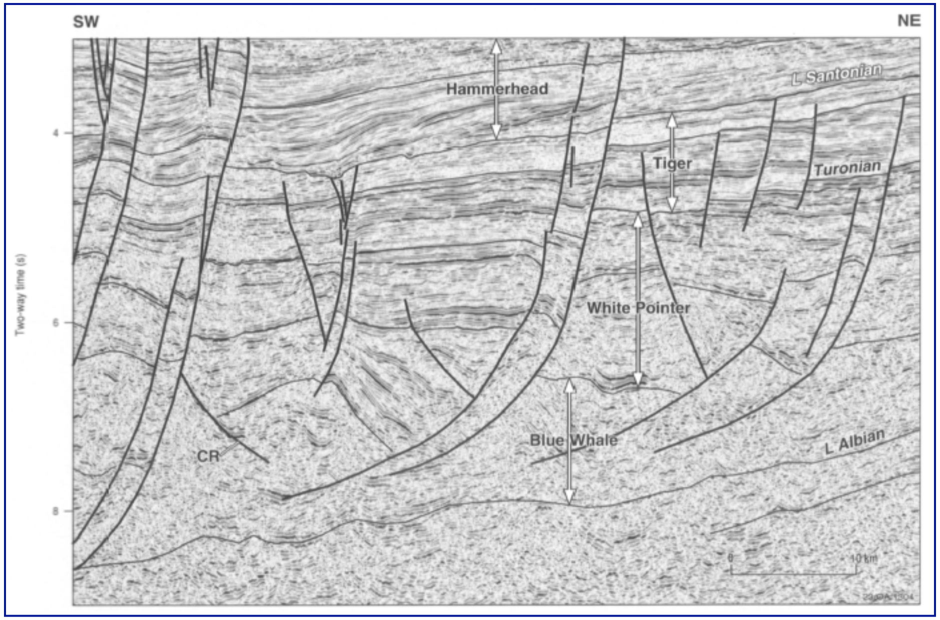


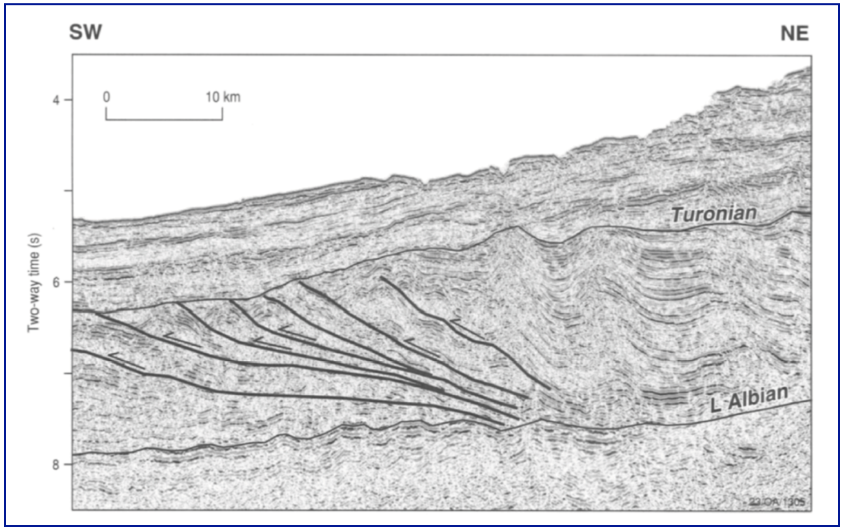
Geoseismic profile across the northern Ceduna Sub-basin. Jumssic-E~ly Cretaceous supersequences abbreviated as follows: SL Sea Lion; M Minke; SR Southern Right; BrW Bronze Whaler

Totterdell and Krassay , 2003

Rapid sedimentation rates in river-dominated environments



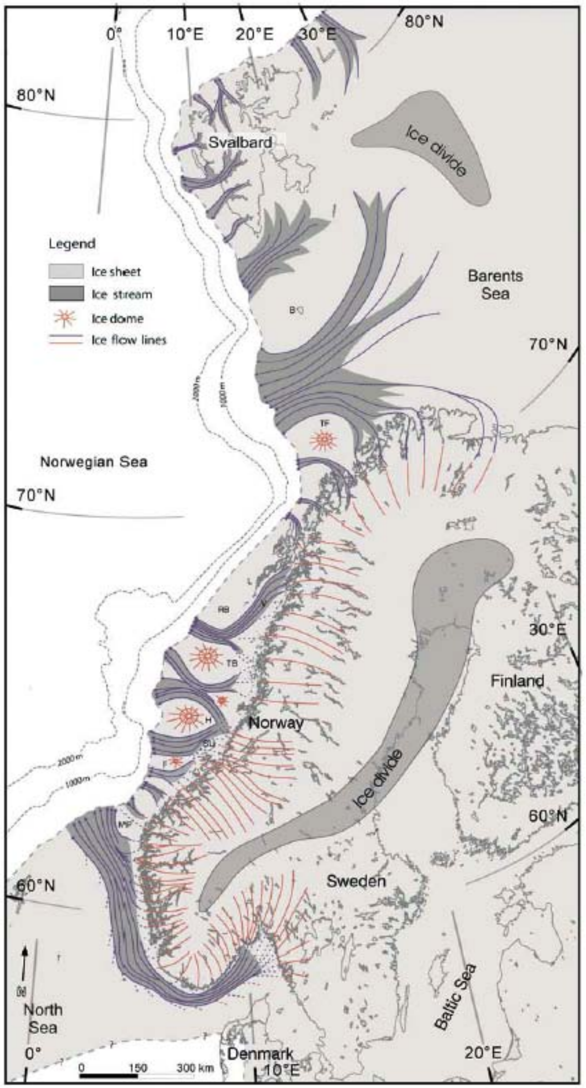


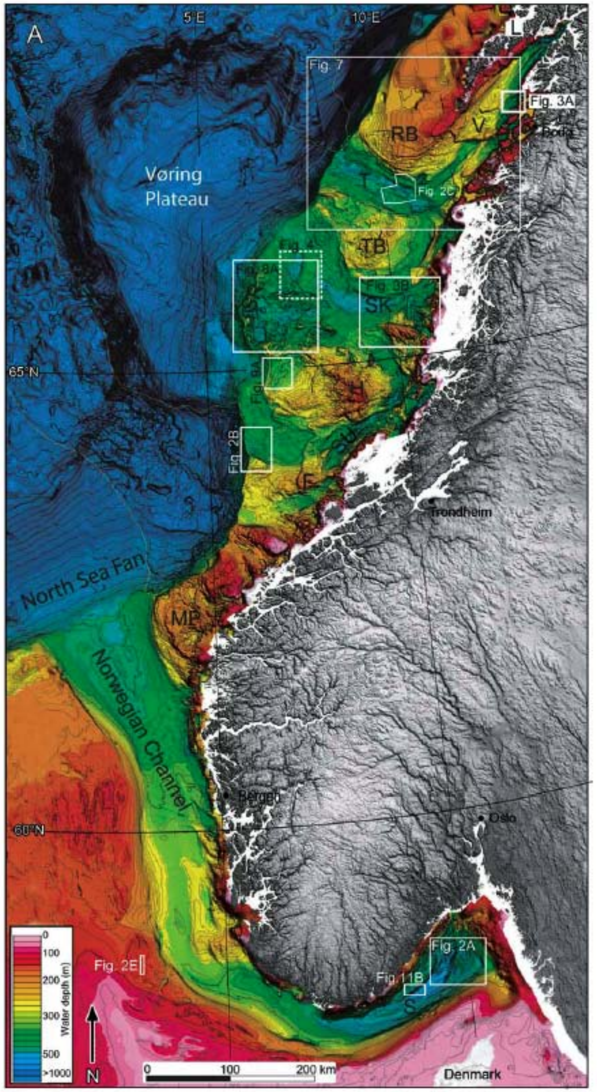


Totterdell and Krassay , 2003

Rapid sedimentation rates in glacial environments

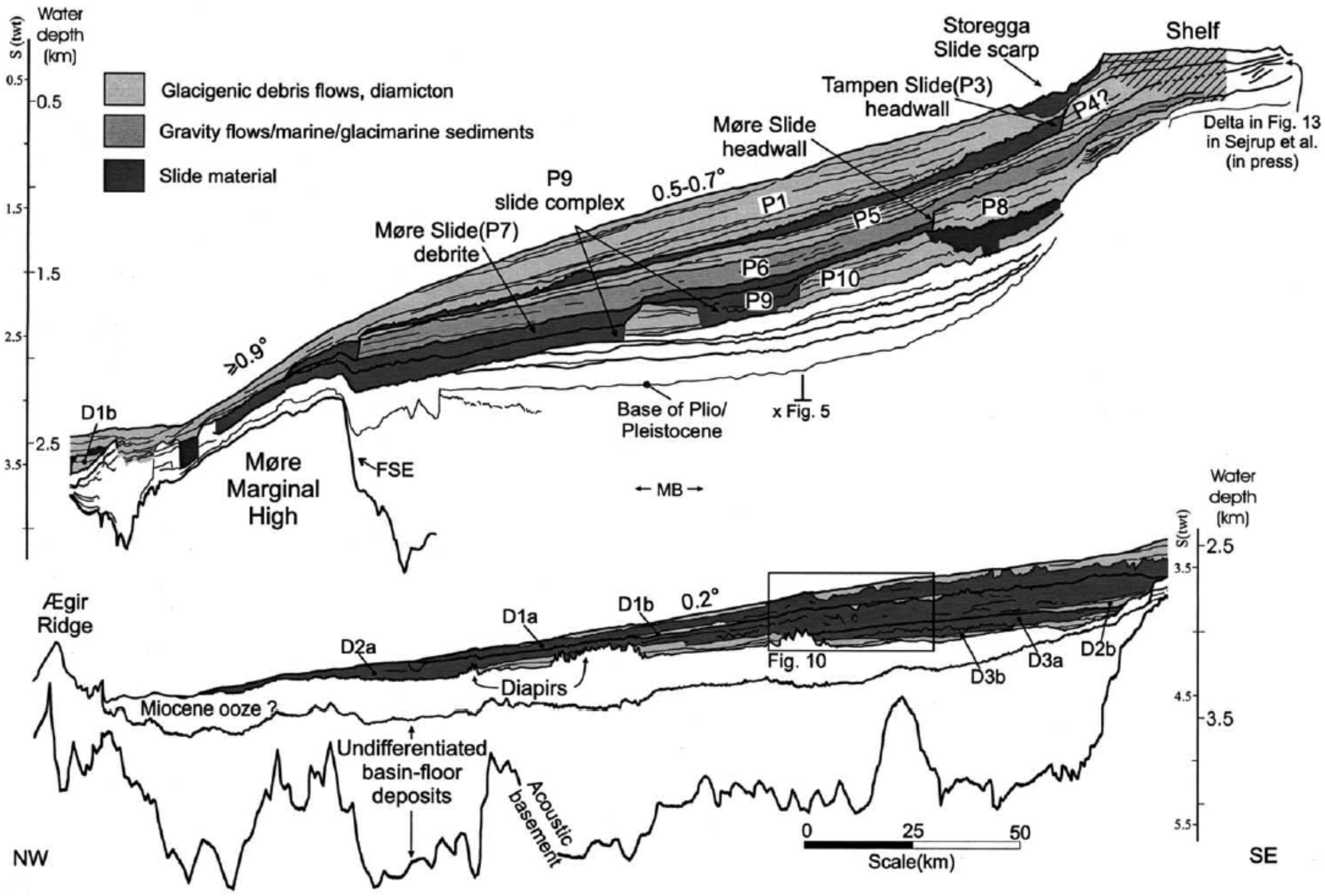
North Sea Fan





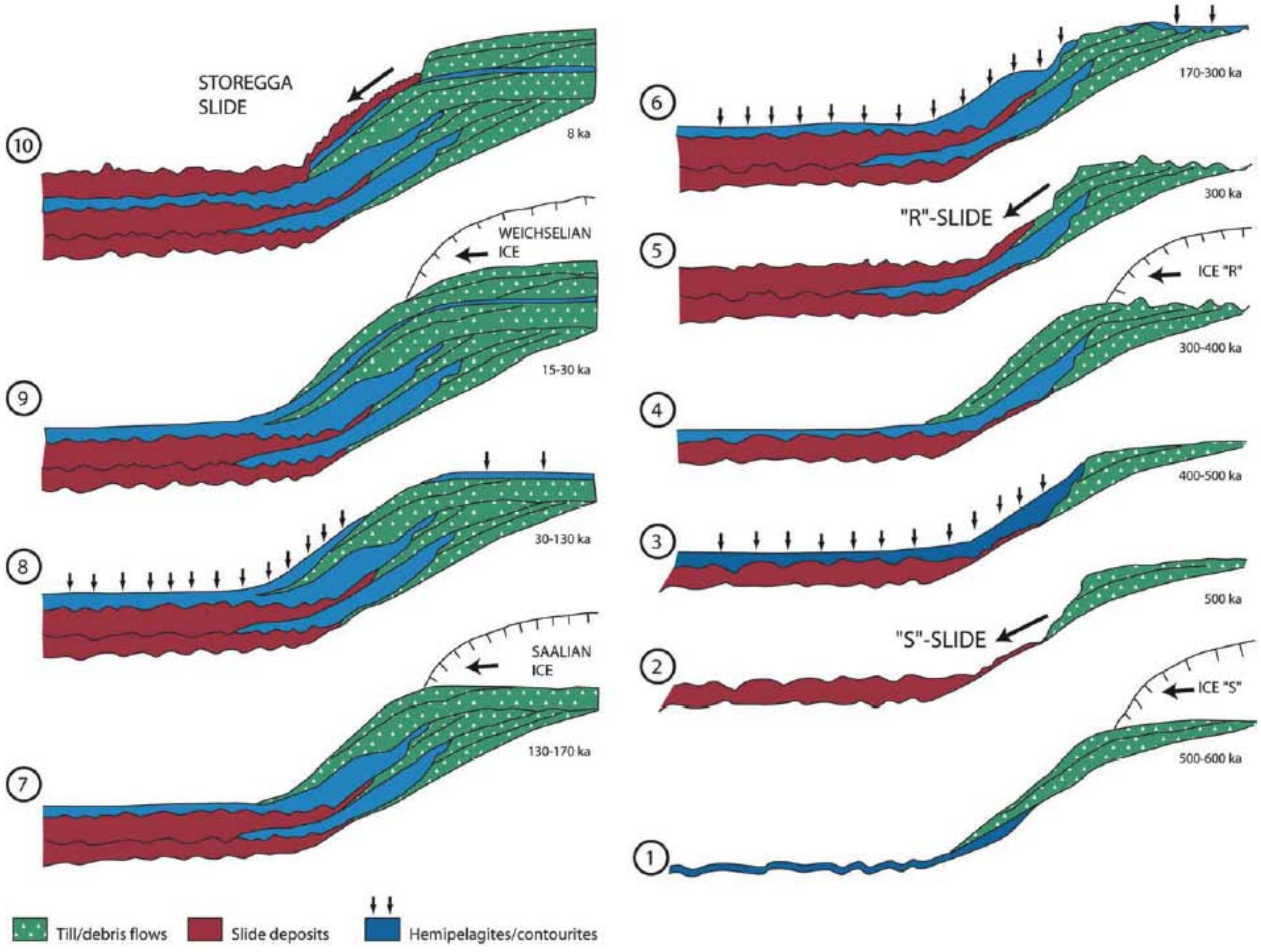
Ottesen et al., 2005

Rapid sedimentation rates in glacial environments



Nygård et al., 2005

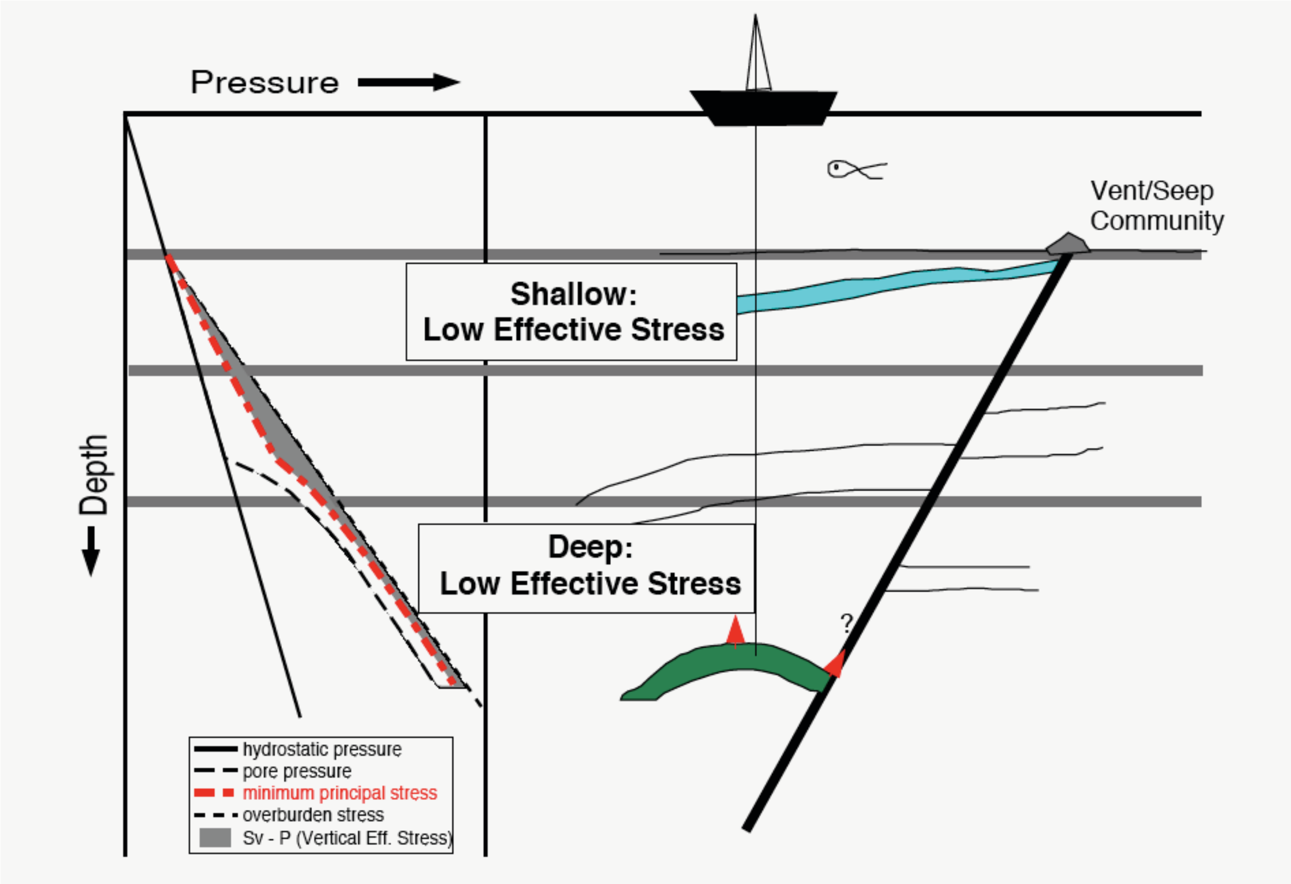
Rapid sedimentation rates in glacial environments



Rise et al., 2005

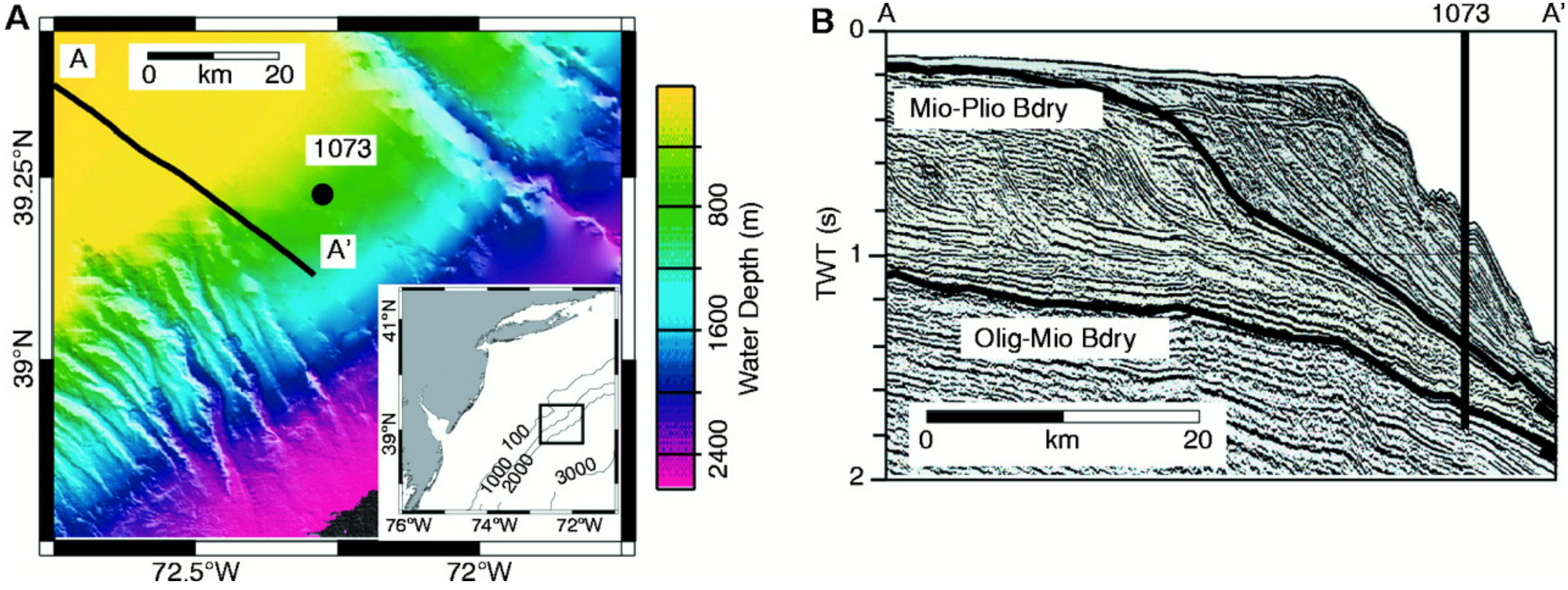
Rapid sedimentation rates

Rapid sediment loading **(>1 mm a^{-1})** is documented as a source of overpressure (*P**, pressure in excess of hydrostatic) in basins around the world. When low-permeability sediments are rapidly loaded, pore fluids cannot escape, and the fluids bear some of the overlying sediment load. In this situation a pore pressure exceeding the hydrostatic pressure develops.



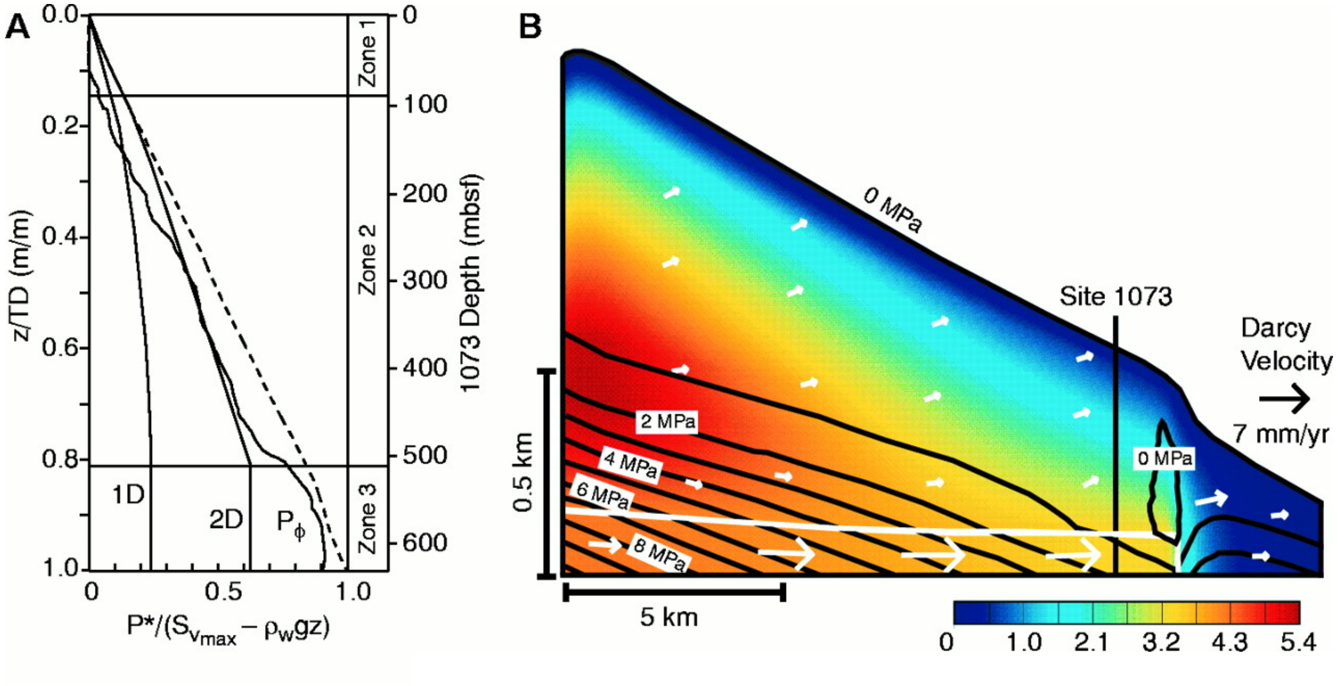
Rapid sedimentation rates

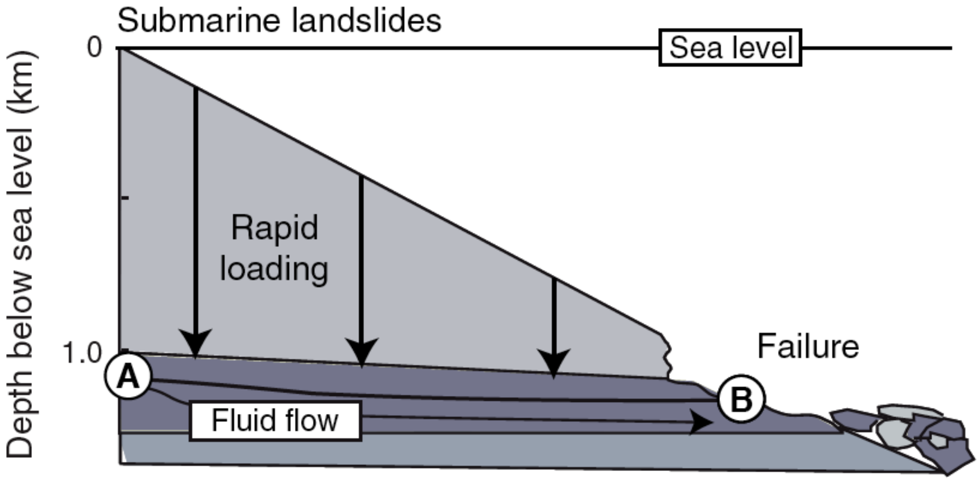
New Jersey Margin.
Rapid sedimentation of clays over a high permeability sand layer



Dugan et al. 2000

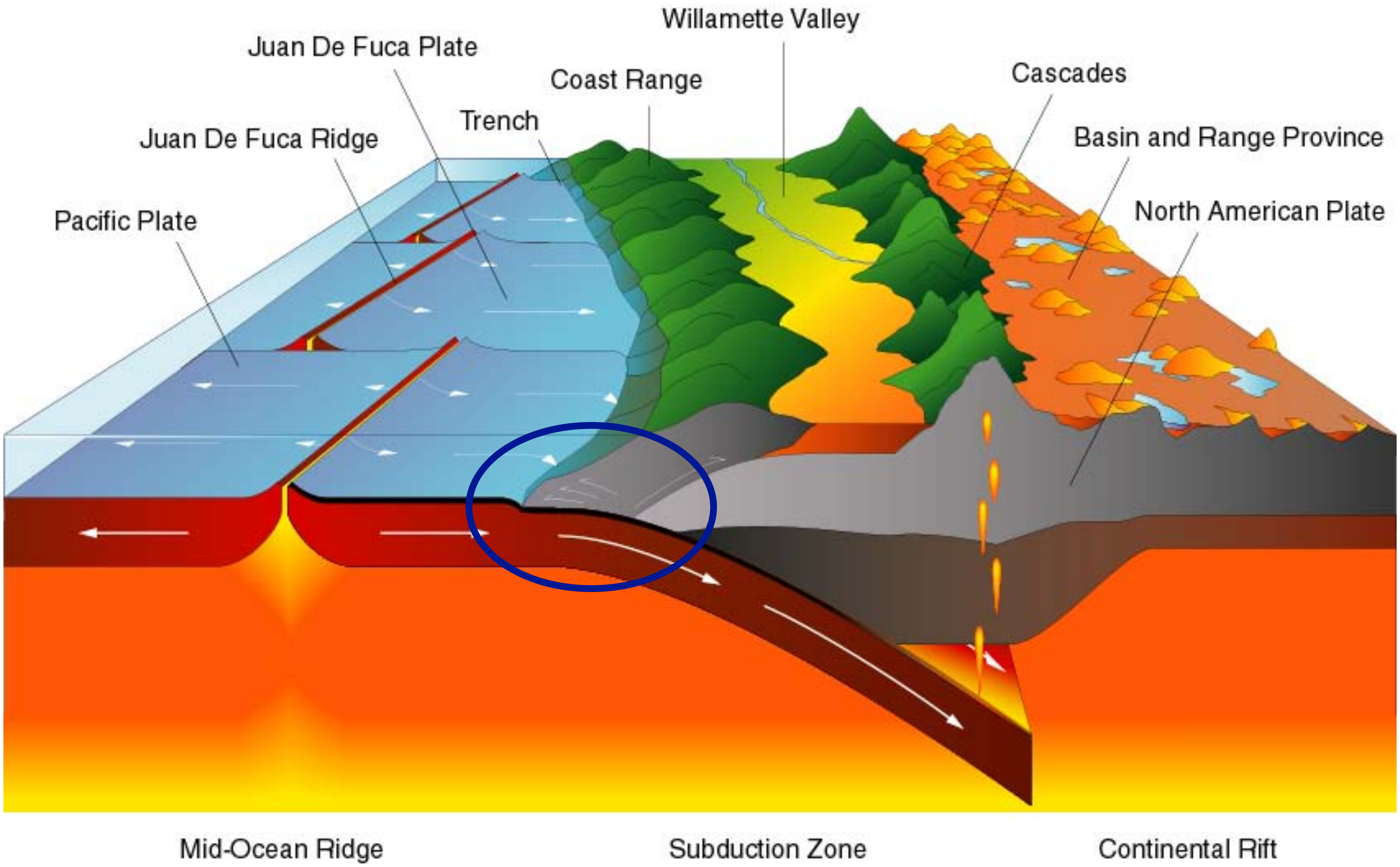
Rapid sedimentation rates



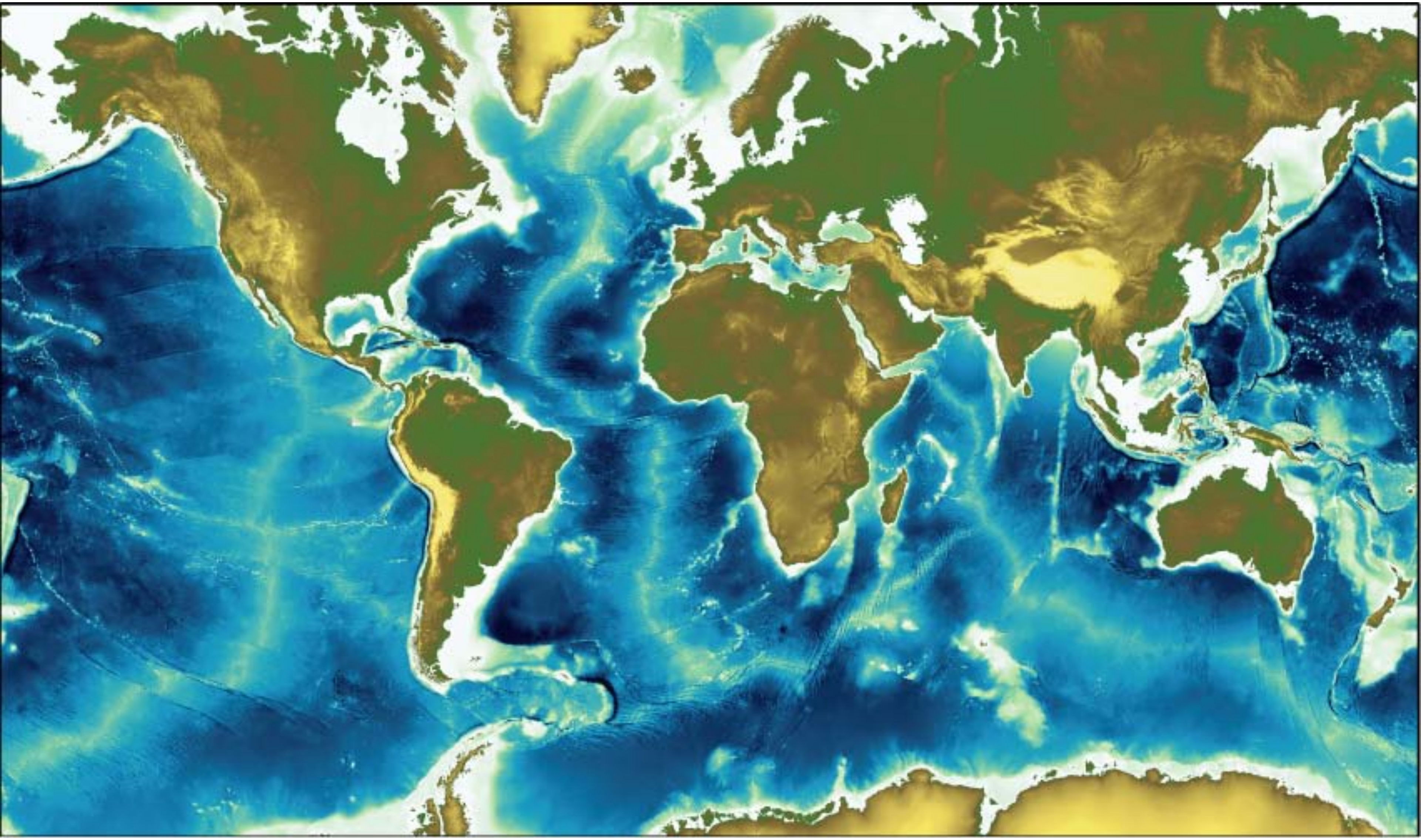


Dugan et al. 2000

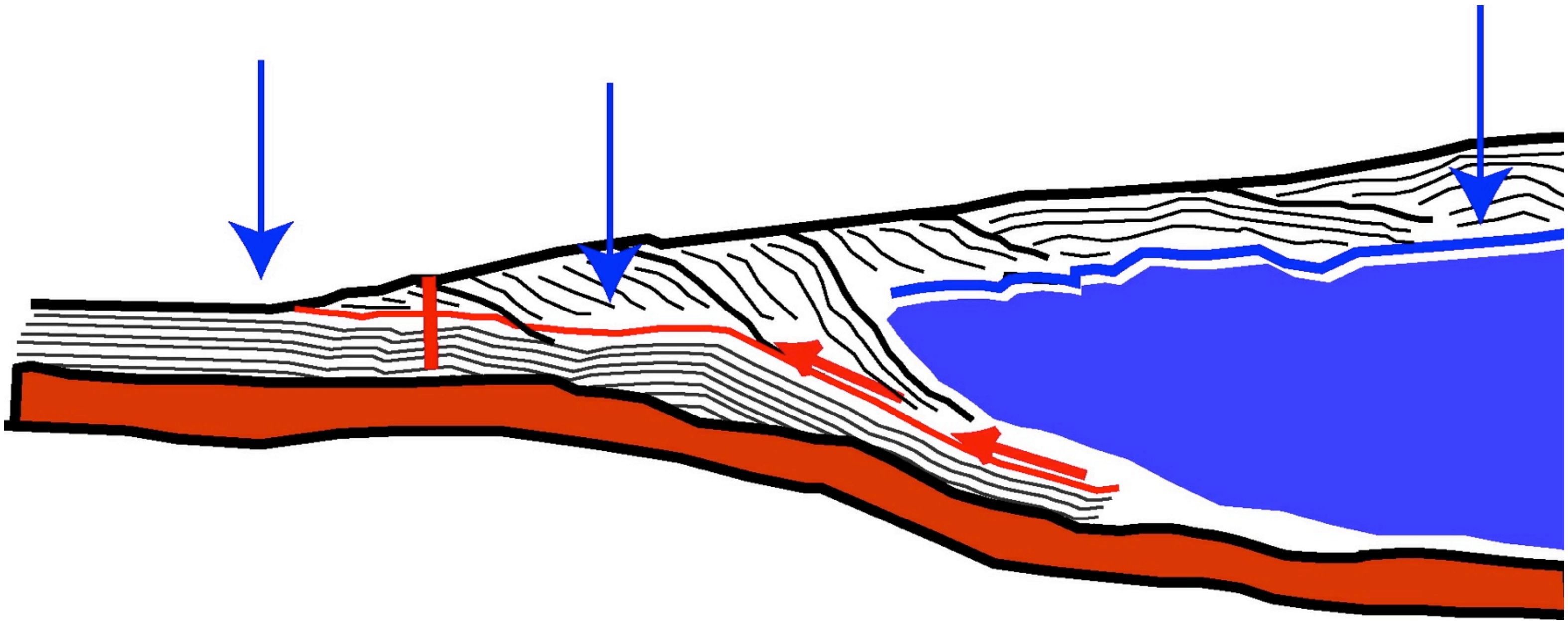
Tectonic stress



Subduction zones are most prevalent in the Pacific Ocean



Tectonic stress



Frontal thrust

Deformed wedge

Slope apron

Tectonic stress

DISPERSED FLOW

If the rate of pore fluid flow is sufficiently low that advection can be accommodated by intergranular permeability, fluids might be expected to leak out of the prism over large areas, as "dispersed flow," driven by some regional sublithostatic pressure gradient.

FOCUSSED FLOW

In contrast, observations establish that flow is often focused at vent sites, confined to fault zones , or expelled from mud volcanoes.

Fluid loss evolves from dispersed to focused flow over time. Observations indicate **episodic flow**, which may be coupled to episodic fault displacement and ultimately to the earthquake cycle.

Tectonic stress

At the deformation front the incoming sediments above the decollement are folded and/or faulted, usually into elongated ridges that rise 200-700 m above the adjacent deep-sea floor (**proto-thrust zone**)

Sequential faulting thickens the accreting sediment section and causes the deformation front to step seaward periodically, widening the lower continental slope. The **thrust faults** rise from the decollement, and controls the amount of sediment that is initially subducted.

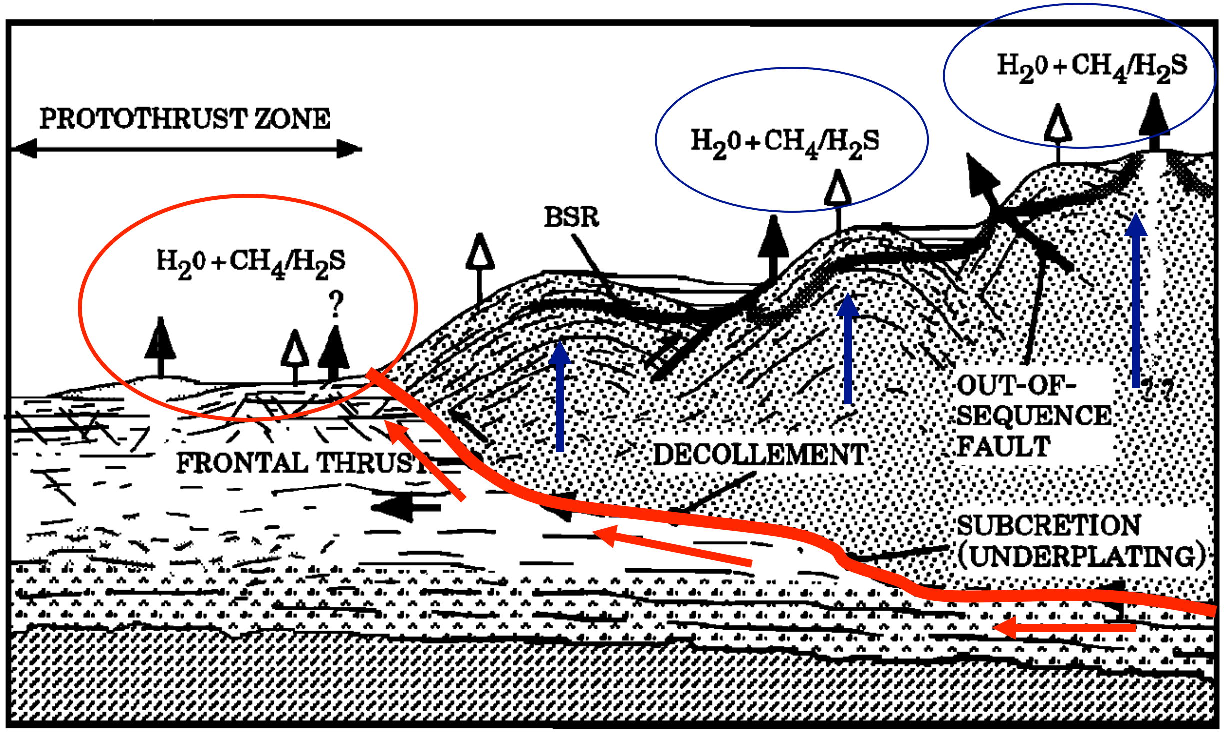
The initial **frontal thrust fault** and others that originate after uplift and folding ("**out-of-sequence" thrusts**) rapidly emplace older, compacted sediments above younger, less consolidated deposits.

This tectonic burial creates overpressures that lead to fluid expulsion.

Tectonic stress

The fluids, are expelled along the decollement are primarily sourced from the underthrust sediments,

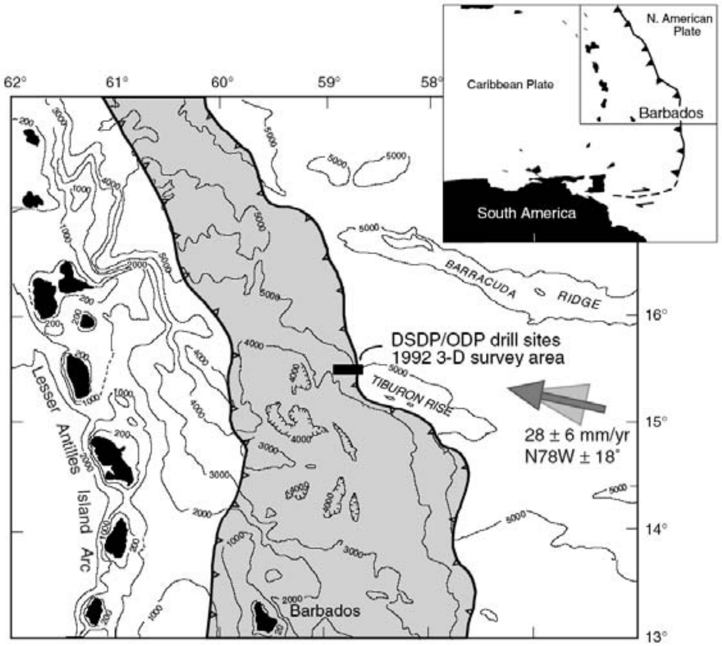
The fluids derived from compaction within the prism generally move upsection to the seafloor.



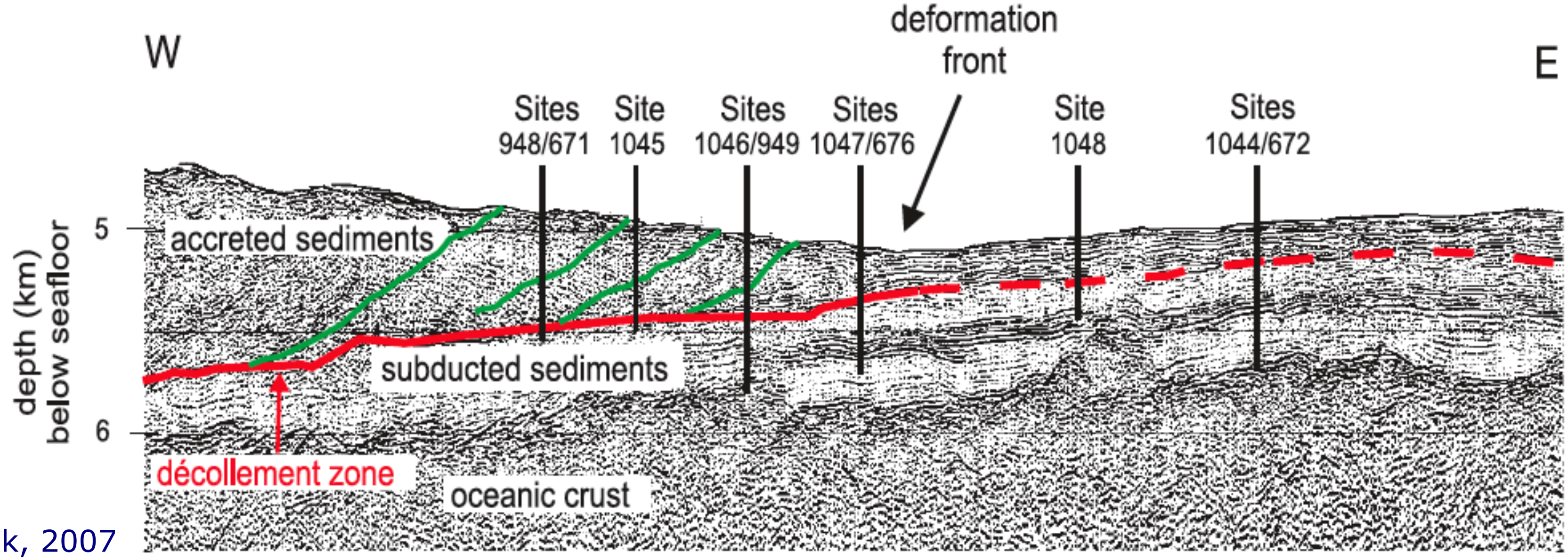
Fluid pressures at the decollement can be approximately lithostatic

Tectonic stress

ODP Leg 171A. Barbados Accretionary Prism



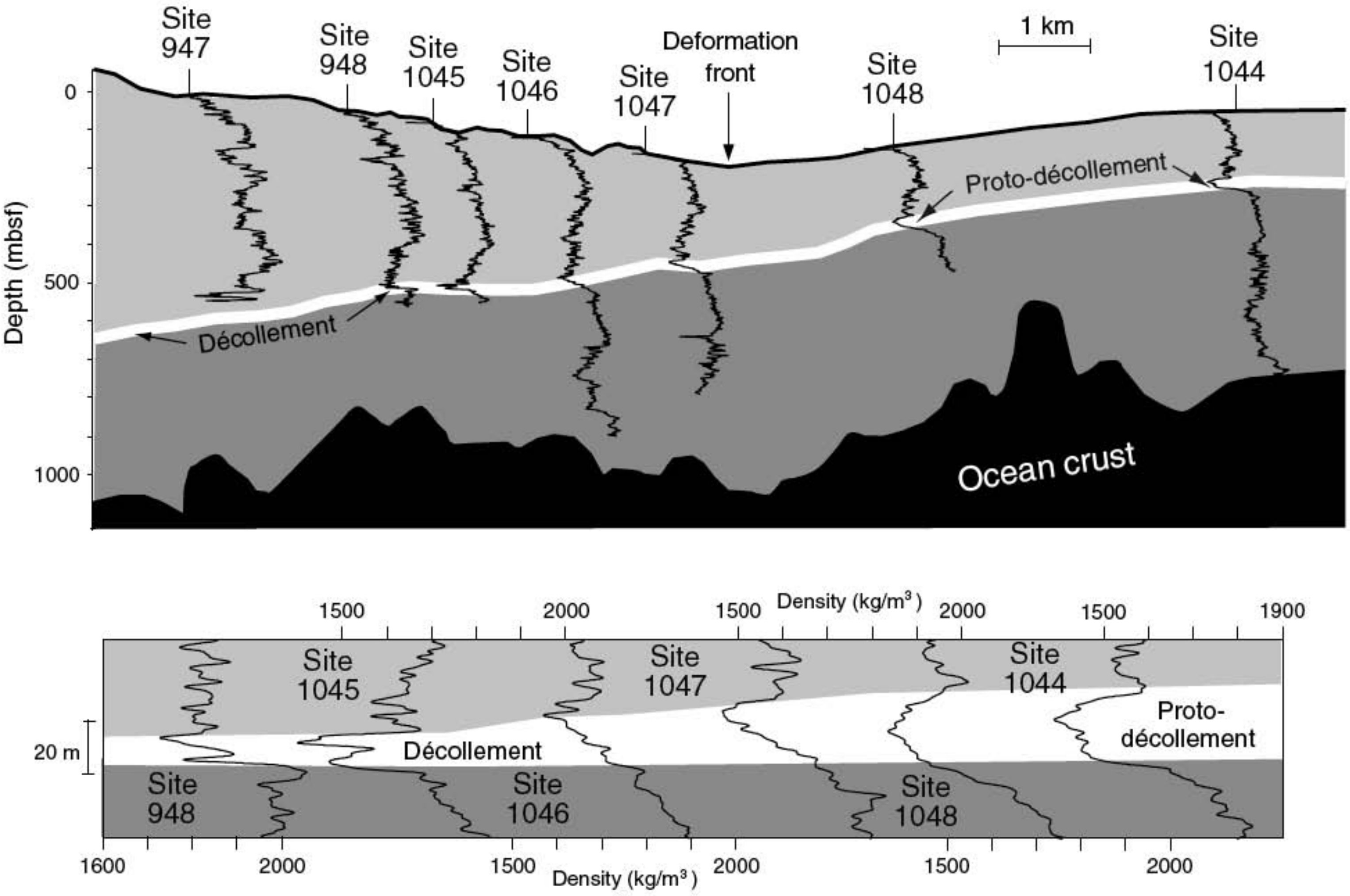
Moore, 2000



Hunze&Wonik, 2007

Tectonic stress

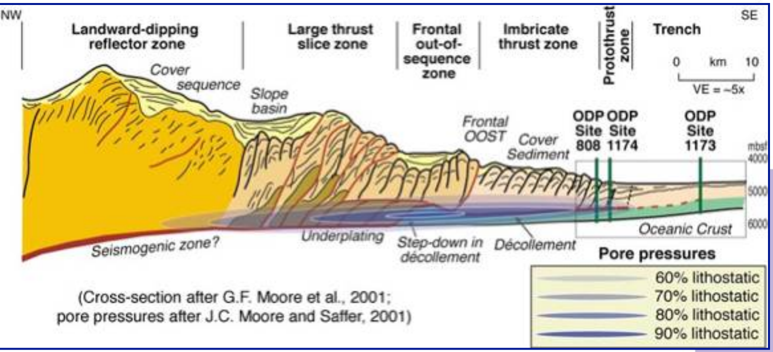
ODP Leg 171A. Barbados Accretionary Prism



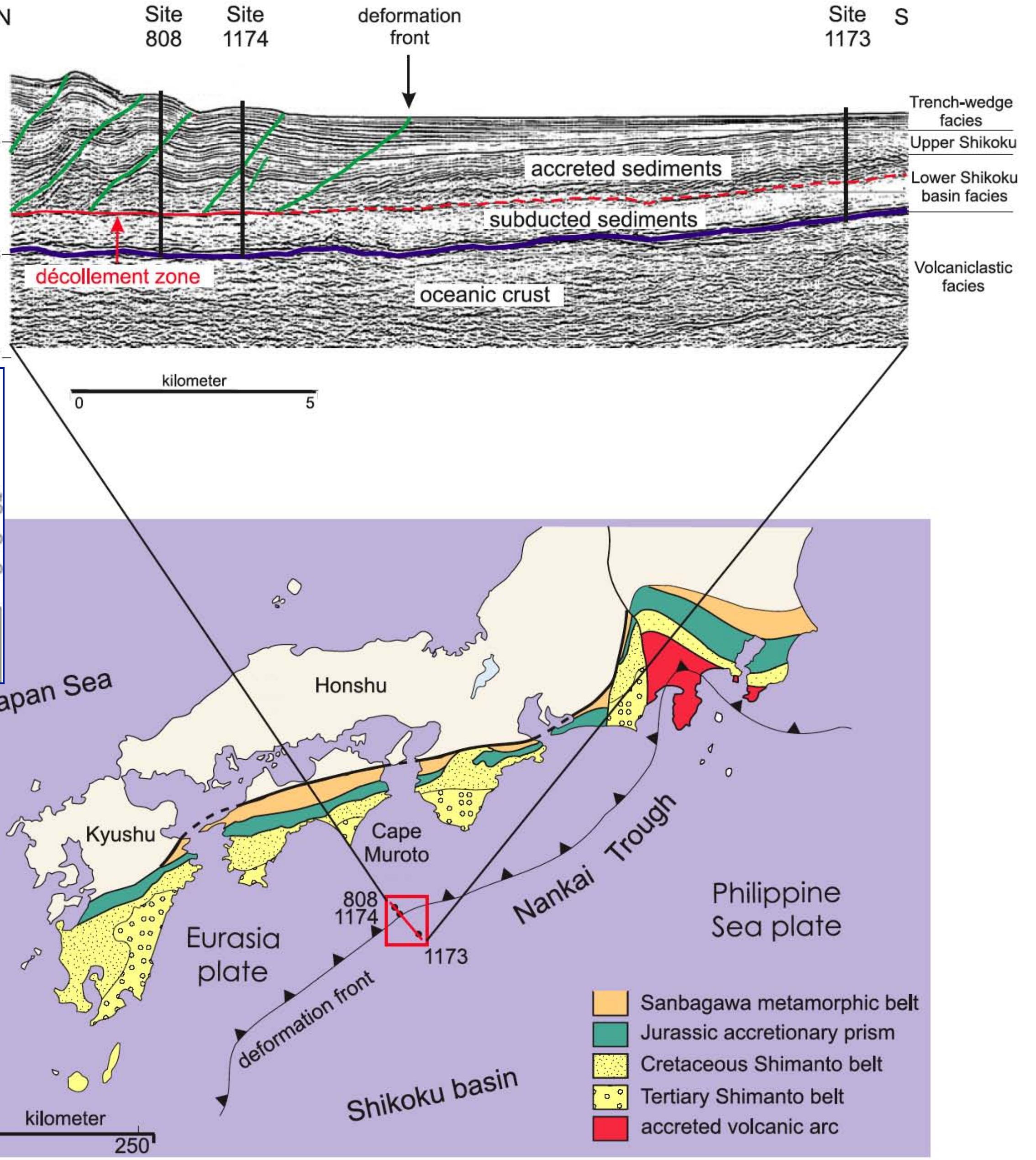
Moore, 2000.

Tectonic stress

ODP Legs 131, 190, and 196. Nankai Accretionary Prism



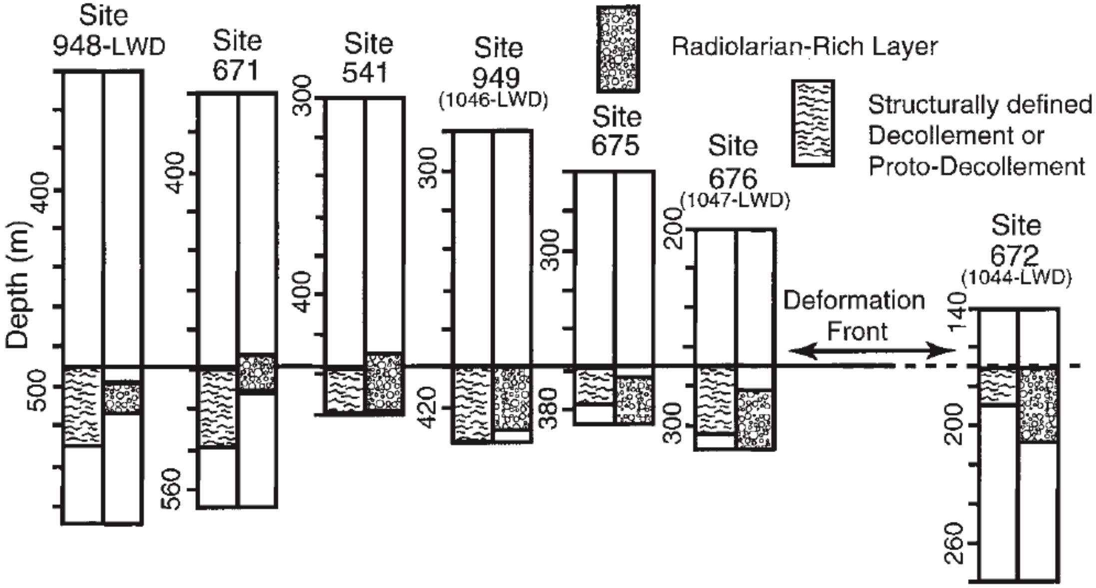
(Cross-section after G.F. Moore et al., 2001; pore pressures after J.C. Moore and Saffer, 2001)

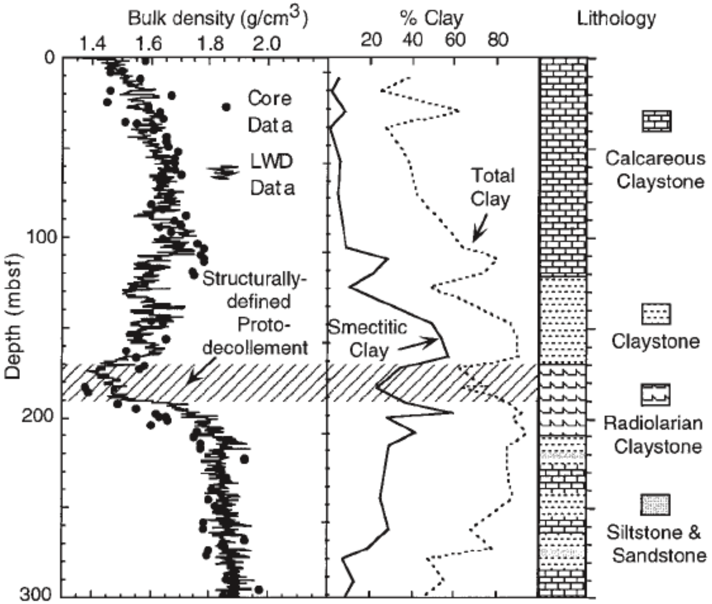


Hunze&Wonik, 2007

Tectonic stress

ODP Leg 171A. Barbados Accretionary Prism





The decollement zone initiates in a low-density smectitic radiolarian claystone.

During underthrusting, the decollement zone consolidates heterogeneously due to shear-induced collapse of the radiolarian tests and clay fabric. This consolidation locally increases the pore pressure and facilitates underthrusting.

More et al., 1998

Factors that determine excess pore water pressure and influence the consolidation state

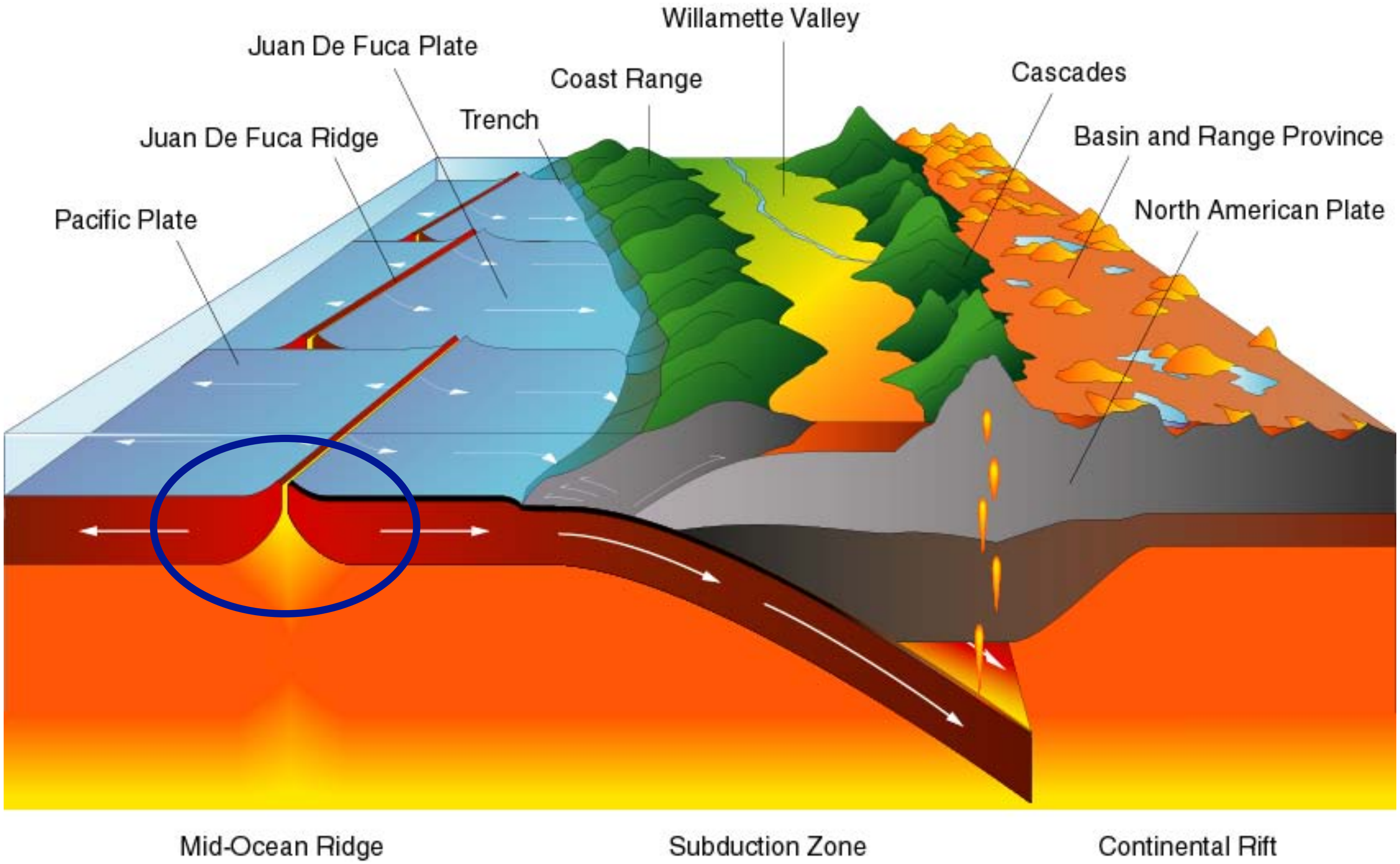
The most common factors are:

- **Stress-related factors**
    - *Rapid sedimentation rate*
    - *Tectonic stress*
- **Fluid volume increase mechanisms**
    - Temperature increase
    - Mineral transformation
    - Hydrocarbon Generation
- **Fluid movement mechanisms**
    - Osmosis
    - Hydraulic head
    - Hydrocarbon buoyancy

(**Swarbrick and Osborne, 1998**. Bryant et al., 1974; Sangrey, 1977; Arthur et al., 1980; Demaison & Moore, 1980; Bryant et al.,1981).

Temperature increase

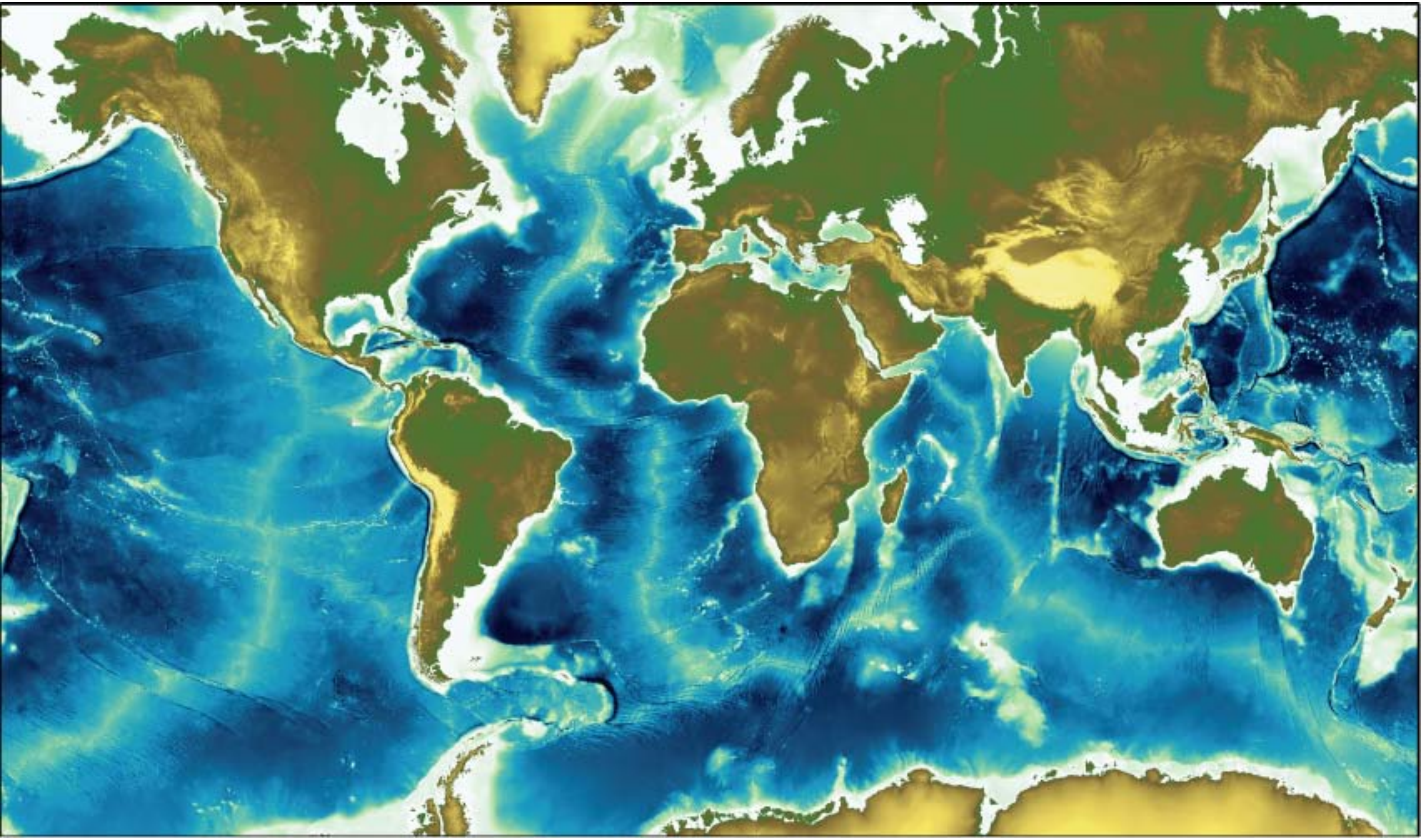
Heat transfer at mid-ocean ridges



Mid ocean ridges (spreading centres) are present in all oceans. The spreading rate is fast in the Pacific, and slow in the Atlantic

Temperature increase

Heat transfer at mid-ocean ridges



Aquathermal Expansion of water

Compressibility: is a measure of the relative volume change of a fluid or solid as a response to a pressure (or mean stress) change.

Compressibility: $$\beta = -\frac{1}{V}\frac{\partial V}{\partial p}$$ where V is volume and p is pressure

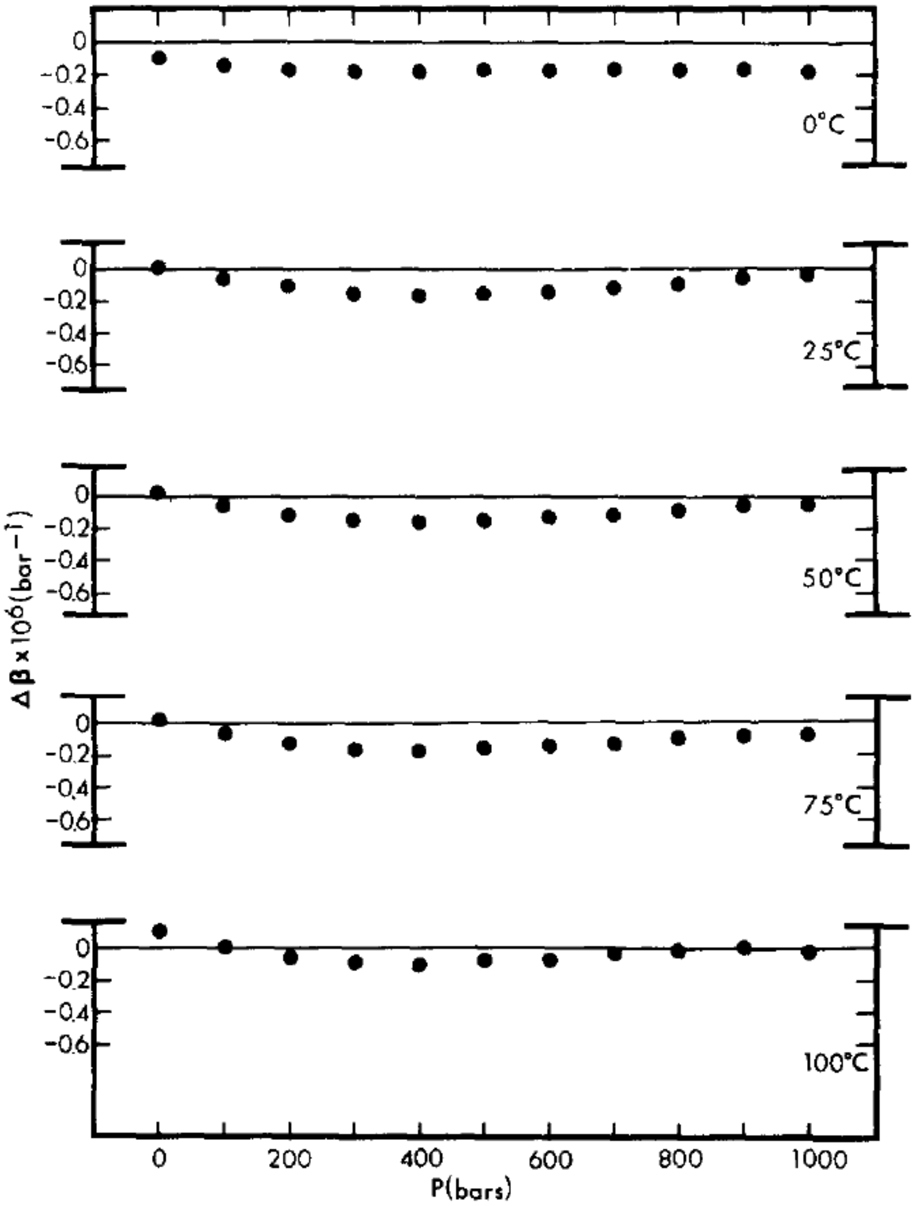
At 0 °C the compressibility is **$5.1 \cdot 10^{-5}$ bar^{-1}**
As the pressure is increased the compressibility decreases, being **$3.9 \cdot 10^{-5}$ bar^{-1}** at **0 °C** and **1000** bar.

The low compressibility of water leads to them often being assumed as **incompressible**. The low compressibility of water means that even in the deep oceans at 4000 m depth, where pressures are $4 \cdot 10^{7}$ Pa, there is only a 1.8% decrease in volume.

Fine and Millero, 1973

Temperature increase

Compressibility of water



Fine and Millero, 1973

Temperature increase

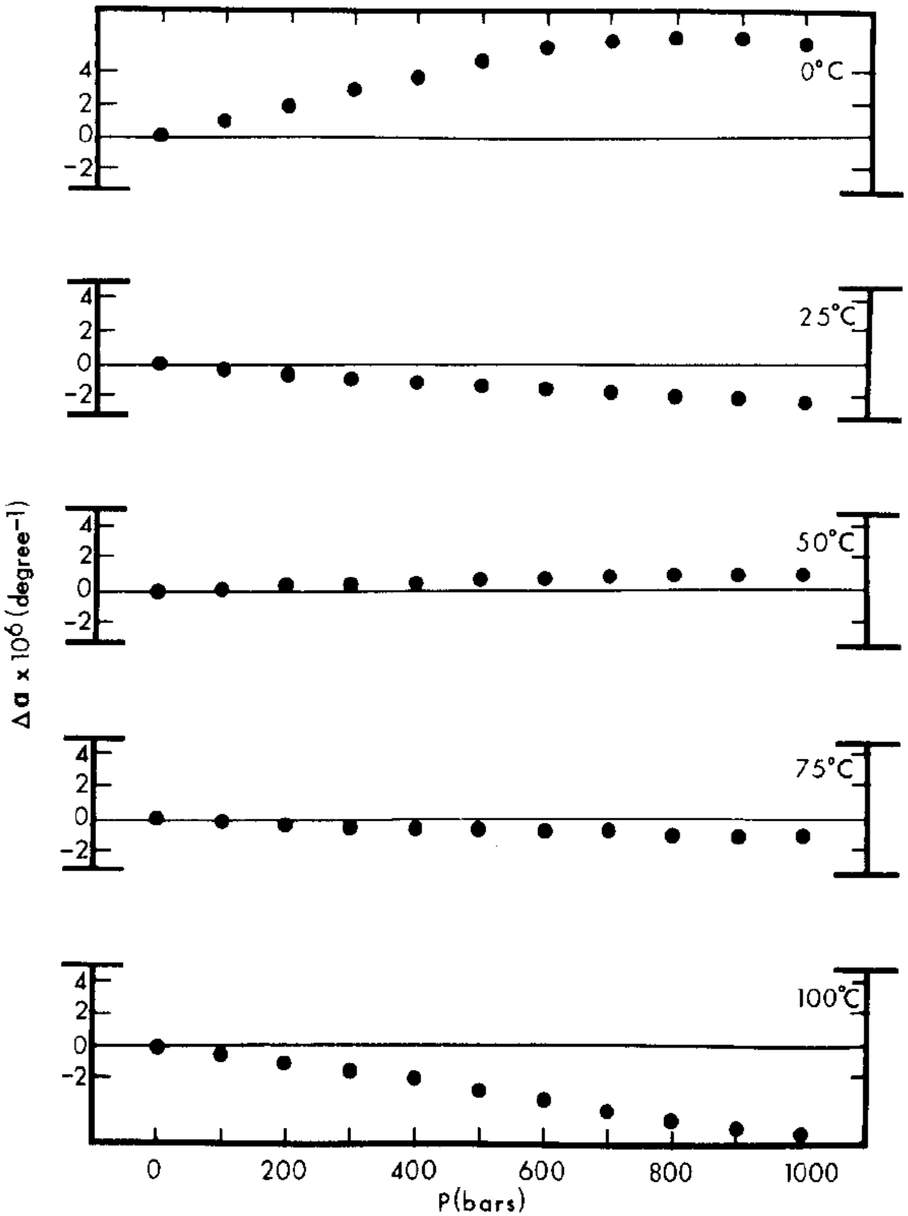
Aquathermal Expansion of water

Thermal expansion:

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

where V is volume and T is temperature

Typical value $5 \cdot 10^{-4}$ °C^{-1}



Fine and Millero, 1973

Temperature increase

If a body of water is **SEALED** in a vessel, the pressure will increase rapidly if temperature rises above +4ºC.

E.g. Pressure increase of 55MPa from heating sealed water from 54 to 93ºC (caused by a volume increase of 1.65%!)

The problem is that in nature, **absolute sealing of water in geological formations is practically impossible**. In addition, the reduction of viscosity of water during heating, facilitates its escape through pores and fractures even at low permeability.

Numerical modelling shows little overpressure in muds with extremely low permeability ($3 \cdot 10^{-14}$ mD)

Only cementation can provide a seal that is nearly absolute.

Swarbrick, and Osborne, 1998

Temperature increase

Sub-surface boiling and supercritical water

In deep sedimentary basins with total depths beyond 10 km, it is conceivable that pore waters can locally (and perhaps temporarily) achieve temperatures of **400 °C** or more.
(assume an average geothermal gradient of 40ºC km^{-1})

When the pressure is too great for water to boil **(>221 bar** for pure water, and **>300 bar** for normal seawater), it attains the supercritical phase, which is neither steam nor liquid, but something in between.
(assume an hydrostatic gradient of 100 bar km^{-1})

Near the critical point on the boiling curve, the density of water changes rapidly, and is intermediate between that of liquid water (1 g cm^{-3}) and low-pressure water vapour (<0.001 g cm^{-3}).

At typical supercritical conditions, the density is approximately **0.3 g cm^{-3}**

Hovland et al., 2006

Temperature increase

Sub-surface boiling and supercritical water

Hydrogen bonding is reduced in supercritical water (non-polar fluid)

Supercritcal water can be regarded as a non-polar fluid that is able to **dissolve organic liquids (oils)**, but **unable to dissolve common sea salts.**

Supercritcal water is also highly **corrosive to silicate rocks**, a property that is of particular importance when studying deep hydrothermal systems and alteration of rocks and sediments.

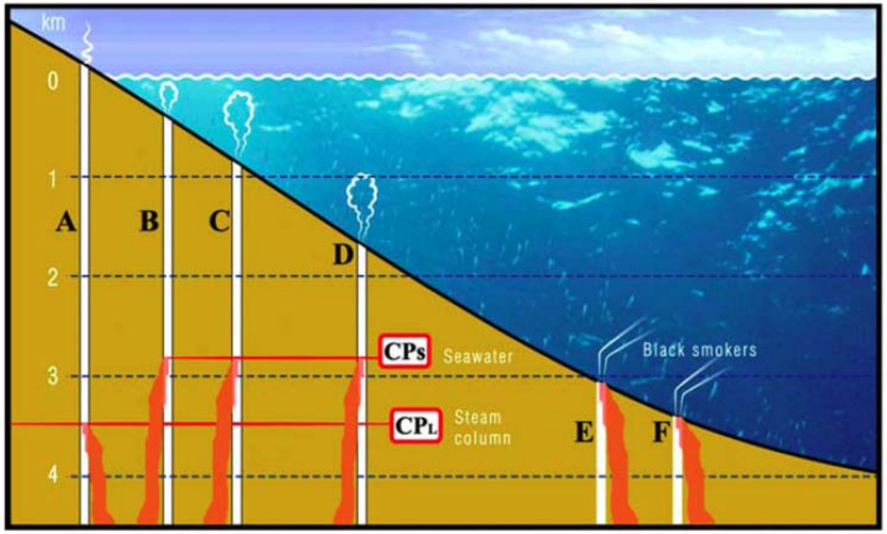
Martin Hovland proposes that many deeply rooted piercement structures, like mud volcano, mud diapirs, asphalt volcanoes, can be triggered by supercritical water movement in rocks.

Hovland et al., 2006

Temperature increase

Geological conditions for supercritical water flow:

1) A deep sedimentary basin >10 km deep.
2) An underlying crust, which is either relatively thin (hot), or which contains fault intersections with high heat-flow.
3) Excess water either within the crust or the sediments, which becomes exposed to the anomalous heat sources (at fault intersections).
4) Local anomalously high or increasing temperature gradients with depth.



CP = Critical Point

Column A shows the association between boiling fresh water and depth on land. The steam column forms above a deep-lying heat source (>374ºC) and the CP occurs deeper than 3 km, because the density of the steam column is low.

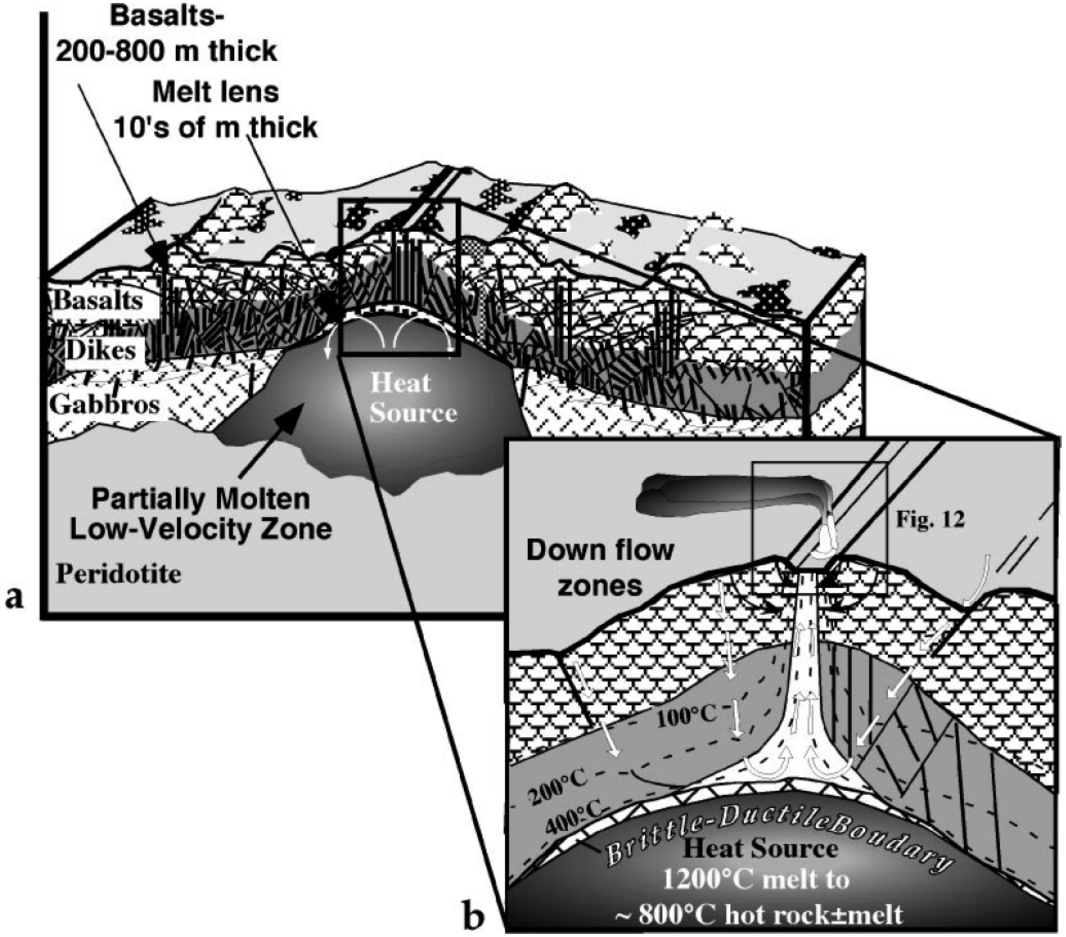
In the ocean, beneath a cold seawater column, above a deep-lying heat source (>405ºC), the CP lies at around 2800m (pressure of about 300 bar). Colums B, C, D have a CP lying at this depth, even though the seafloor depth varies.

The two hot vents E, F have CPs at different depths, as they are both deeper than 2800m and the CP is defined by temperature (cooling by seawater) rather than by depth. These ‘black smokers’ invariably construct sulphide and silicate chimney structures formed by mineral precipitation as supercritical fluids are instantly cooled by seawater.

Hovland et al., 2006

Temperature increase

Heat transfer at mid-ocean ridges



Mantle-crust relationships beneath ridge crests.

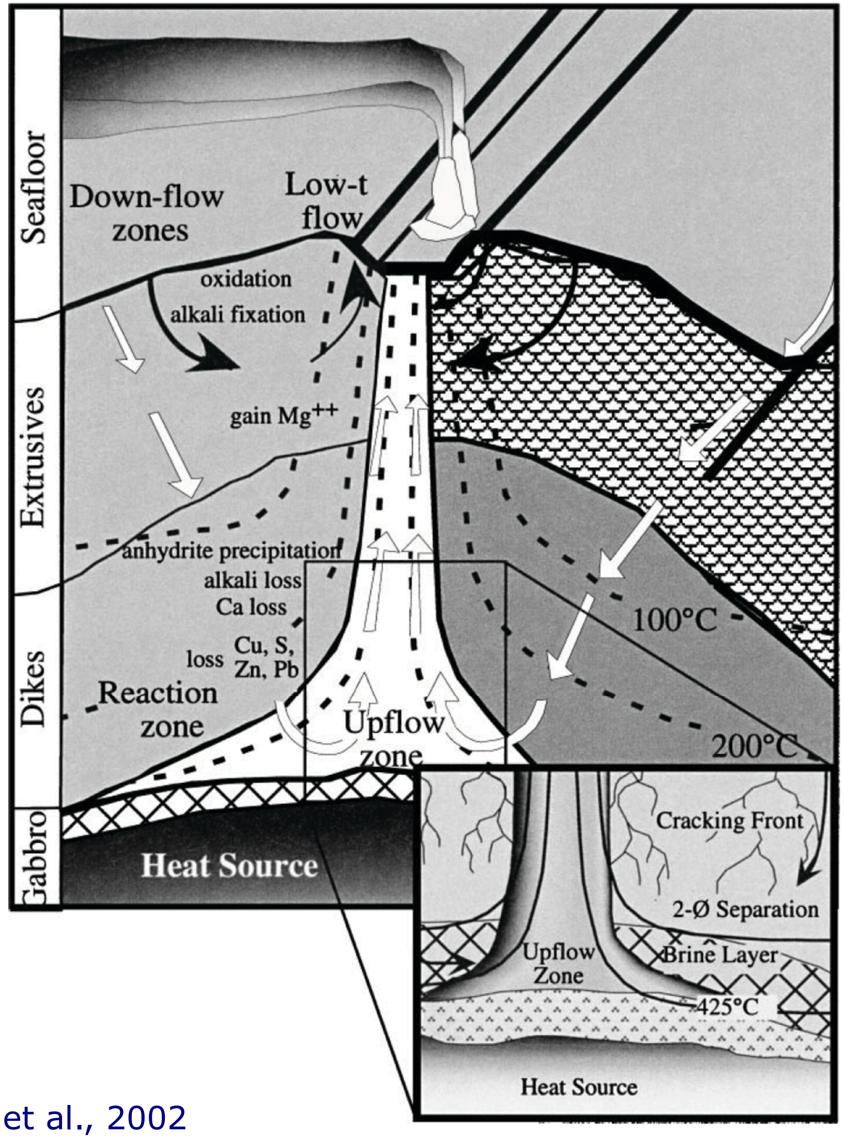
(a) Crustal magma chambers, fed from melt percolating through the underlying mantle section, typically form at depths of 1–4 km below the seafloor.

(*b*) Steep thermal gradients resulting from intrusion of 1200±C basaltic melt into cool, water-saturated and porous crustal rocks drive hydrothermal circulation beneath the spreading centers. Hightemperature limbs of the resultant hydrothermal cells focus metal-rich, acidic fluids onto the seafloor, which form sulfide deposits upon mixing with cold, oxygenated seawater.

Kelley et al., 2002

Temperature increase

Heat transfer at mid-ocean ridges



Schematic showing chemical reactions and mineral precipitation associated with down-welling recharge systems, low-temperature shallow circulation, and deep penetration by hydrothermal fluids into the reaction zone.
At the base of many mid-ocean ridge hot springs, the seawater-derived fluids undergo either boiling or condensation.
If condensation occurs for prolonged time periods, a brine layer may develop deep within the crust. These fluids may be expelled during waning of high-temperature hydrothermal flow.

Fluid penetration is believed to occur during downward migration of small fracture networks and cooling of the crust along a cracking front, which allows fluids to have continual access to hot, fresh rock.

Kelley et al., 2002

Mineral transformation

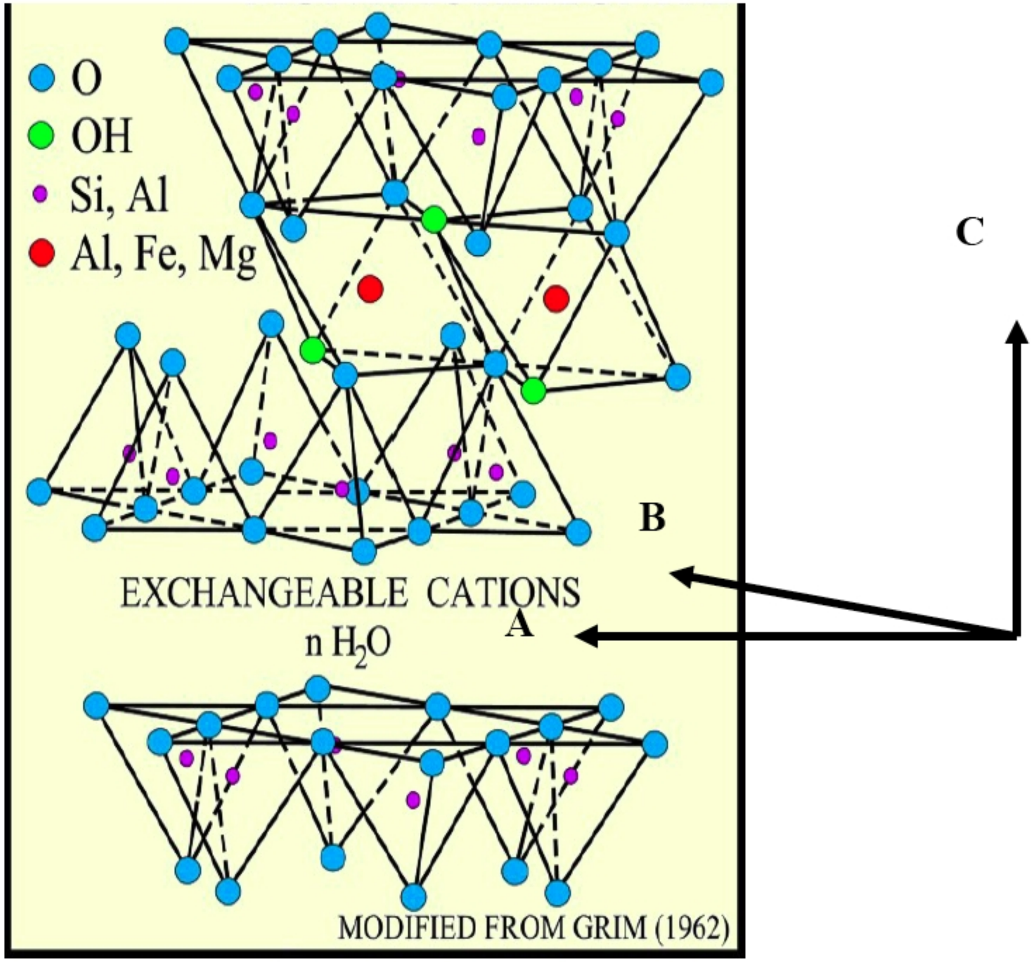
- Smectite dehydratation
- Smectite to Illite Transformation
- Gypsum to Anhydrite Dehydratation

Mineral transformation

- Smectite dehydratation

A significant feature of the smectite group is its very high surface area. These clay platelets are held together by a weak electromagnetic force (Van der Waal's bonds), and there is a considerable amount of area to which up to ten layers of water can bond. The result is a low density "swelling clay”.

Under normal hydrostatic conditions smectite contains 2 or 3 layers of water.



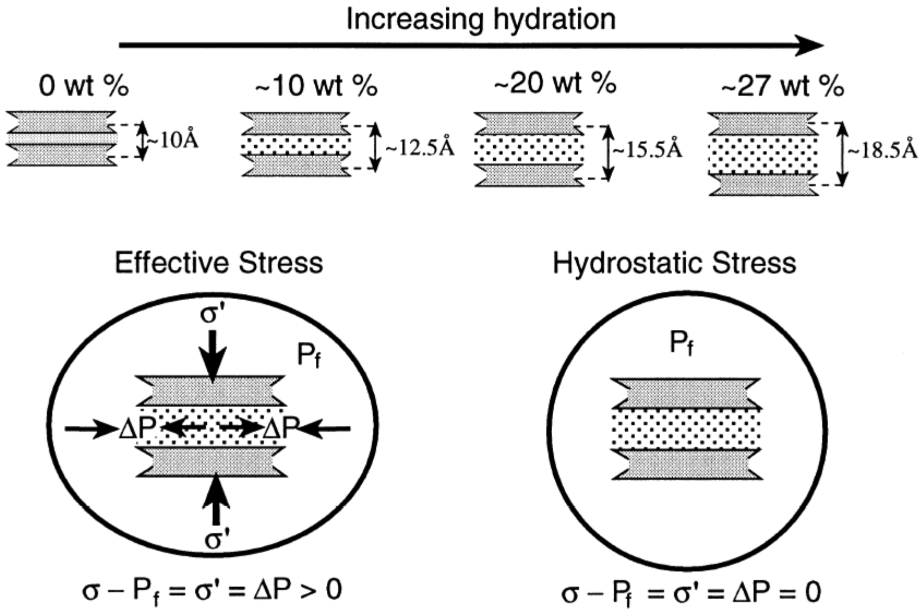
Iain Hillier, Baker Hughes INTEQ

Mineral transformation

- Smectite dehydratation

Smectite changes with burial. Effect of lithostatic stress.

1) Increasing pressure will drive out the loosely bound water



Smectites can partially dehydrate from 18.5 Å to 15.4 Å hydrates when they are subjected to effective mineral framework stresses above 1.3±0.3 MPa
(sort of squeezing of interlayer water out of the mineral structure)

Fitts and Brown et al., 1999

Mineral transformation

• Smectite dehydratation

Role of temperature

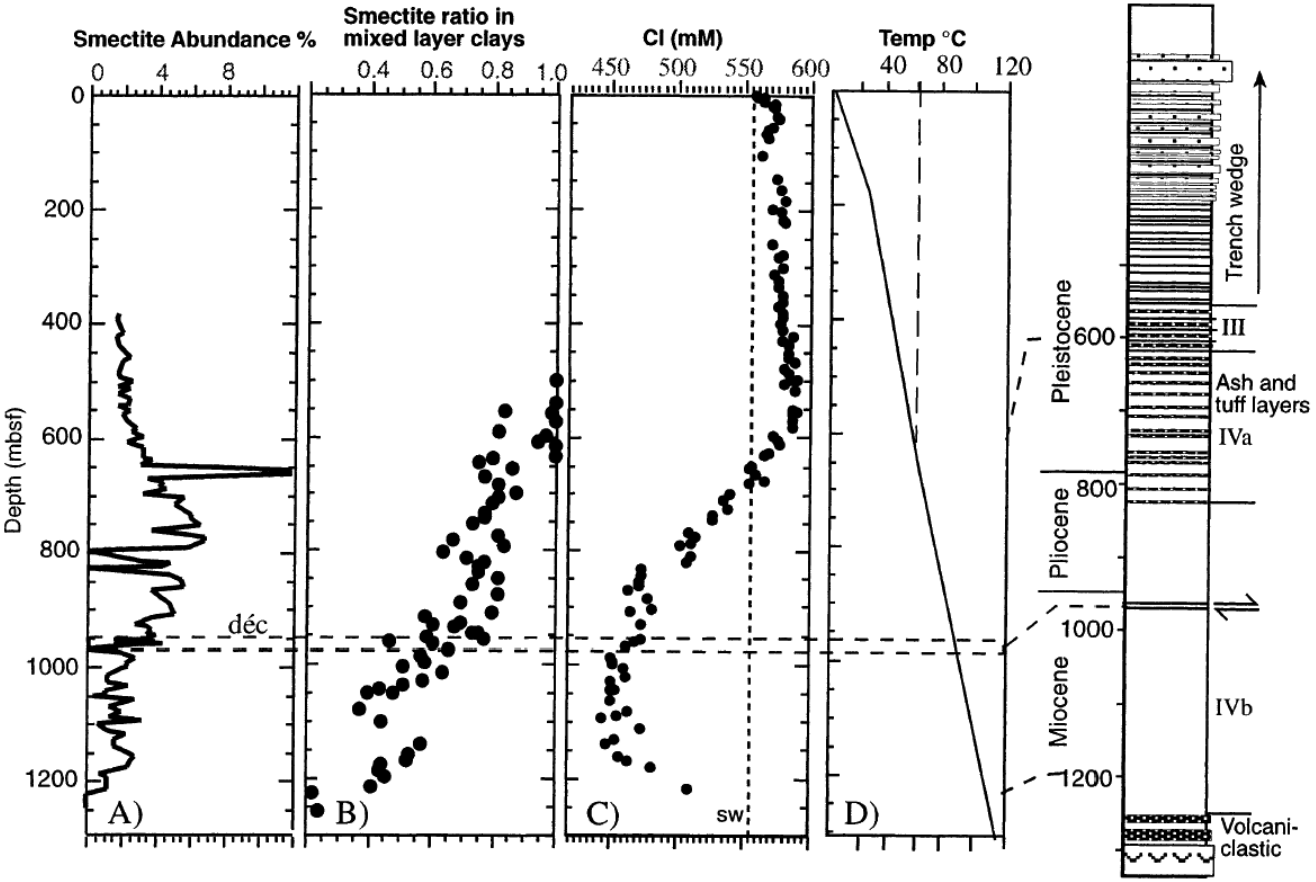
1) Under stress (burial) at temperature of about 60ºC, smectite molecules will lose interlayer water except the last two layers.
2) With temperature of about 100ºC one the penultimate layer will be displaced.
3) A further rise in temperature to 172° - 192° C is required to drive off the last layer, which is very closely bound between the clay plates.

Although the initial dewatering may actually cause some overpressures, the resulting extra hydraulic pressure will also tend to inhibit further dewatering. Therefore, under most conditions the simple dewatering process will not lead to excessive overpressure (**volume increase of about 4%**), since there is a negative feedback loop at work.

Iain Hillier, Baker Hughes INTEQ

Mineral transformation

- Smectite dehydratation



Possibility that the observed low-Cl anomaly is a compound effect of both lateral flow and in situ smectite dehydration.

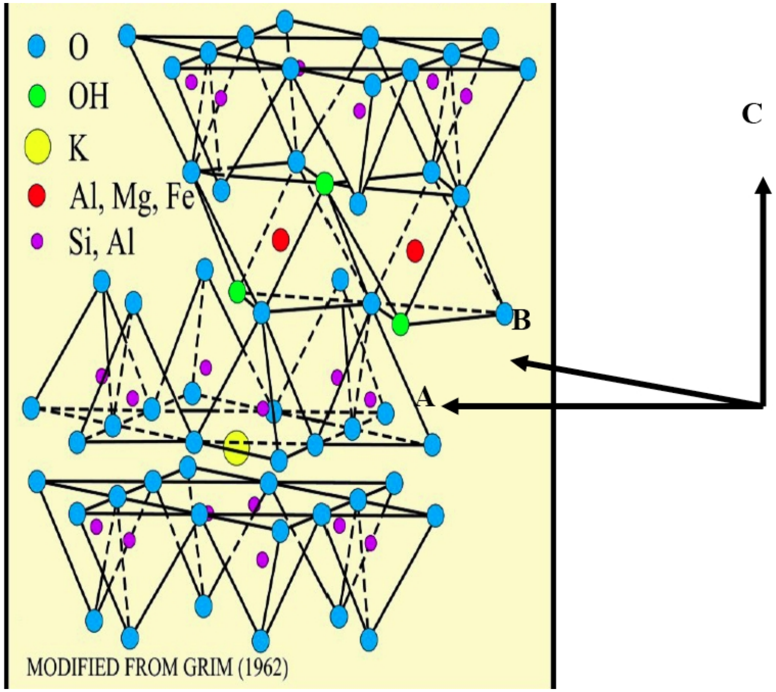
Brown et al., 2001

Mineral transformation

- Smectite to Illite Transformation

During Burial, smectite alters chemically to produce illite

Smectite + (Al, K) - (Na, Ca, Mg, Fe, Si, water) >>>> **illite**



MODIFIED FROM GRIM (1962)

Iain Hillier, Baker Hughes INTEQ

Mineral transformation

- Smectite to Illite Transformation

- The reaction is kinetically controlled.
- It usually happens at temperature range of 70ºC - 150ºC
- It often accompanies the loss of the penultimate water layer of smectites.
- It has been shown that in order for the transformation to occur Potassium **(K) ions are required** in the interstitial fluids. These come from K-Feldspars
- Illites do not have the same capacity to adsorb water as smectities.
- The **threshold temperature** for the loss of the penultimate water layer is roughly the same at which **hydrocarbons are generated**.

Mineral transformation

- Smectite to Illite Transformation

The volumetric change of Smectite to Illite is unknown.

Some models say ΔV = + 23-25 %

It is a fact that overpressure is often detected in wells in correspondence with such diagenetic transformation.

Consequences during the transformation :

- Compressibility increases
- Overpressure
- Disequilibrium compaction
- Perhaps reduction of permeability by silica cementation

Mineral transformation

- Gypsum to Anhydrite Dehydratation

Heating gypsum to between 100 °C and 150 °C partially dehydrates the mineral by driving off approximately 75% of the water contained in its chemical structure.

The temperature and time needed depend on ambient partial pressure of H_2O.

The reaction for the partial dehydration is:

$$CaSO_4 \cdot 2H_2O + \text{heat} >>> CaSO_4 \cdot nH_2O + n\ H_2O$$

Anhydrite: $CaSO_4 \cdot nH_2O$, where n is in the range 0.5 to 0.8.

The temperature of the reaction depends on the physical and geochemical environments

Mineral transformation

- Gypsum to Anhydrite Dehydration

Geological factors:

- Thermal conductivity of the overlying rock
- Basal heat flow
- Porosity
- Sedimentation rate

Geochemical factors:

- Activity of pore water
- Pore fluid pressure

Jowett et al., 1993

Hydrocarbon Generation

Hydrocarbon is an organic compound consisting entirely of hydrogen and carbon.

“**Impure**” hydrocarbons include bonded compounds or impurities of sulphur or nitrogen

- 1 **Saturated hydrocarbons** (**alkanes**) are the most simple of the hydrocarbon species and are composed entirely of single bonds and are saturated with hydrogen; they are the basis of petroleum fuels and are either found as linear or branched species of unlimited number. The general formula for saturated hydrocarbons is C_nH_{2n+2}.
- 2 **Unsaturated hydrocarbons** have one or more double or triple bonds between carbon atoms. Those with one double bond are called alkenes, with the formula C_nH_{2n} (assuming non-cyclic structures). Those containing triple bonds are called alkynes.
- 3 **Cycloalkanes** are hydrocarbons containing one or more carbon rings to which hydrogen atoms are attached. The general formula for a saturated hydrocarbon containing one ring is C_nH_{2n}
- 4 **Aromatic hydrocarbons**, also known as arenes which have at least one aromatic ring

Hydrocarbon Generation

Natural gases (alkanes gases)

METHANE CH_4
ETHANE C_2H_6
PROPANE C_3H_8
BUTANE C_4H_{10}

METHANE: OXYDATION OF ORGANIC MATTER BY ANAEROBIC BACTERIA (BIOGENIC)

C_{2+} GASES: THERMAL MATURATION OF ORGANIC MATTER (THERMOGENIC)

IN NATURAL GAS THERE ARE TWO TYPES OF METHANE:

- **Biogenic Methane**

Biogenic methane is produced as an end product of the metabolism of a diverse group of obligate **anaerobic archaea** (killed by even traces of oxygen), generally known as **methanogens**.

Methanogens live in the shallow geosphere such as **marine and lacustrine sediments** (sub-bottom depth less than 1 km, temperature less than 80°C).

Hydrocarbon Generation

In the absence of light, photosynthesis cannot occur. Respiration makes use oxygen of dissolved in water with microbial mediation.

Organic matter (preferably proteins and carbo-hydrates) is decomposed trough hydrolysis by micro-organisms, that produce sugars and aminoacids that more evolved organisms can use.

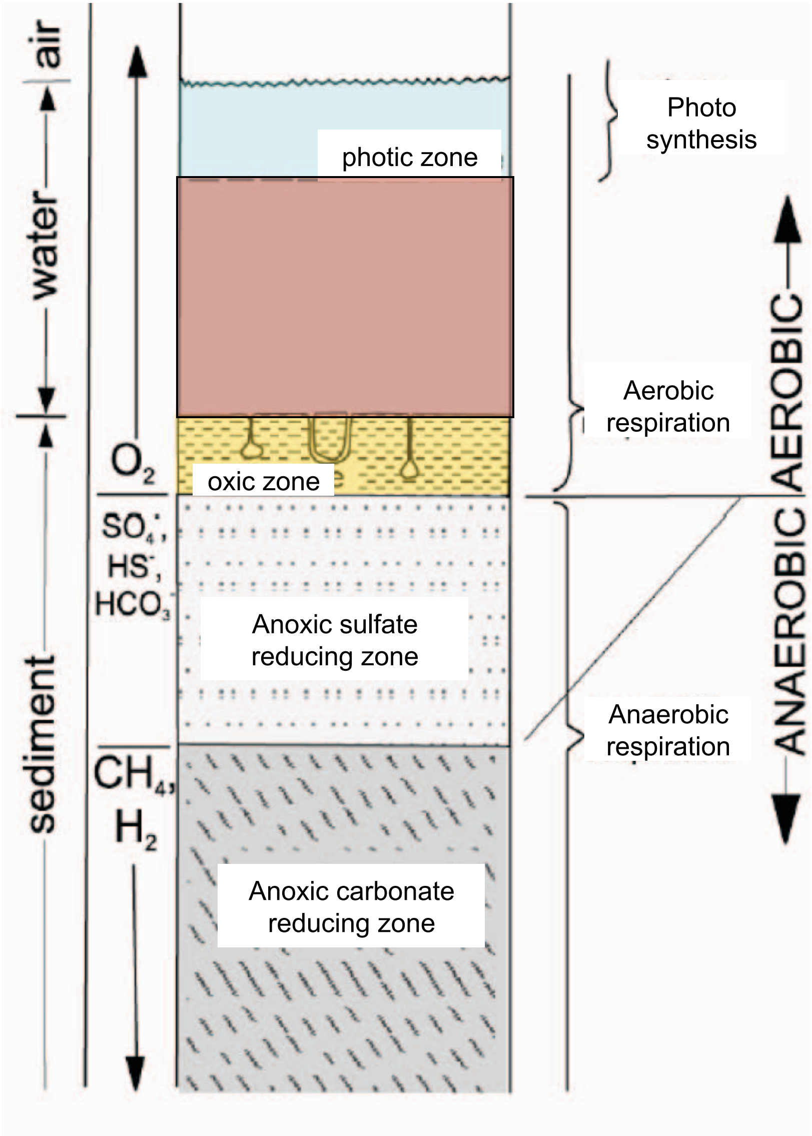
Aerobic respiration in the interstitial fluids of the uppermost sediments can cause **oxygen depletion** in pore water (usually the exchange of fluids between pores and water column is negligible due to low permeability).

Respiration then uses oxygen from Mn-Oxides, Nitrates and Iron-Oxides (**sub oxic Zone**)

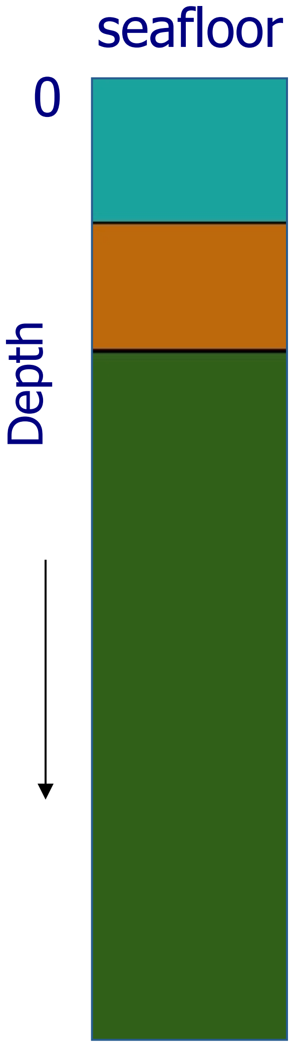
Aerobic respiration is then replaced by **anaerobic fermentation**, that is the microbial reduction of the sulfate ion (SO_4) with release of S^- ions and production of H_2S, and CO_2.

When all the Sulfate ion has been used, the Archea begin to produce methane (CH_4) from the remaining organic matter.

The biogenic production of methane continues until the temperature of 75-80°C is reached.



Hydrocarbon Generation



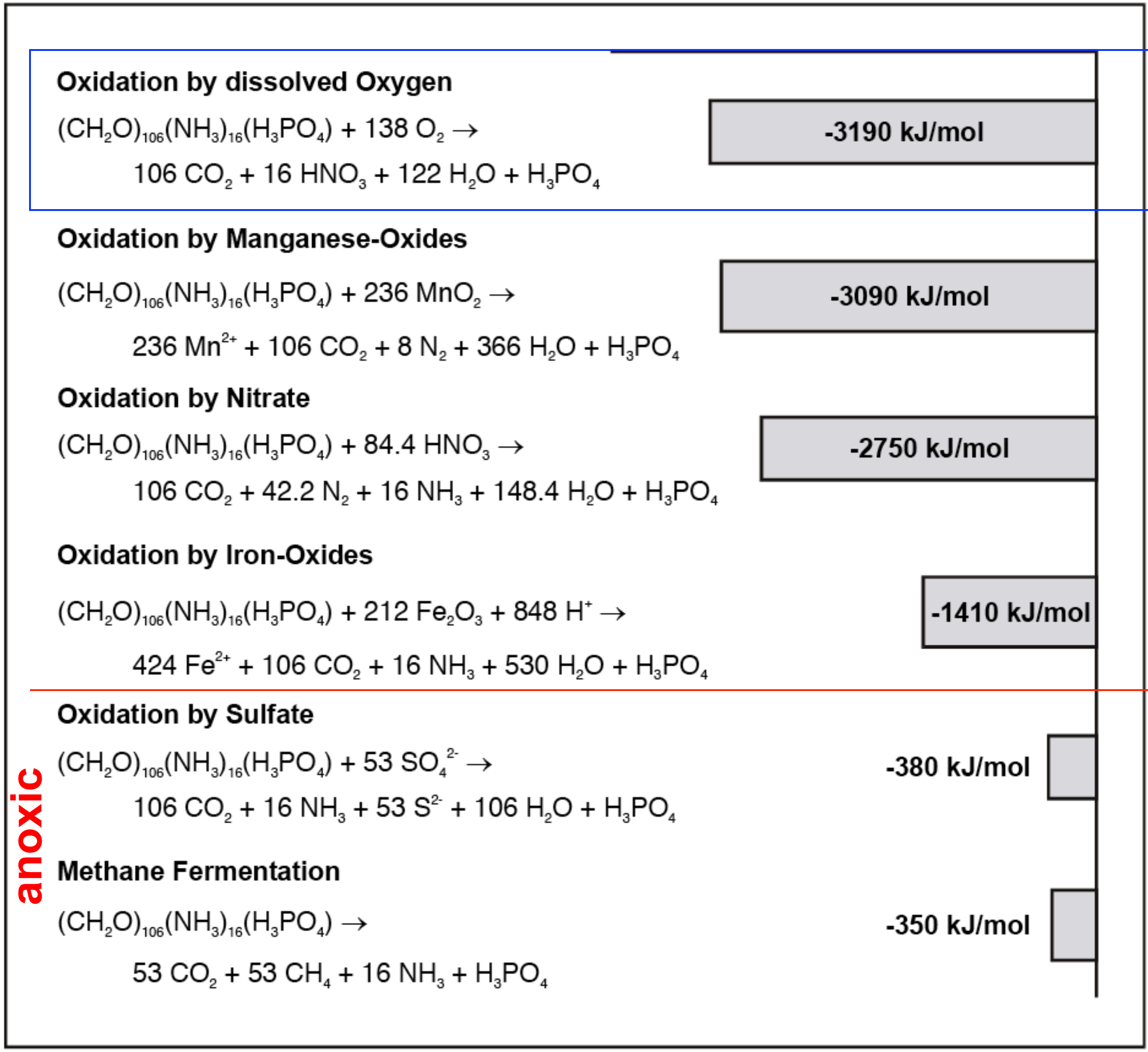
1. Aerobic (respiration):
$CH_2O + O_2 = CO_2 + H_20$

2. Sub-oxic zone

3. Anoxic (sulfate reducing) zone:
$CH_2O + SO_4 = CO_2 + H_2S$

4. Methanogenesis:
$CH_2O = CO_2 + CH_4$

Hydrocarbon Generation



Oxidation by dissolved Oxygen

$(CH_2O)_{106}(NH_3)_{16}(H_3PO_4) + 138\ O_2 \rightarrow$
$106\ CO_2 + 16\ HNO_3 + 122\ H_2O + H_3PO_4$

-3190 kJ/mol

Oxidation by Manganese-Oxides

$(CH_2O)_{106}(NH_3)_{16}(H_3PO_4) + 236\ MnO_2 \rightarrow$
$236\ Mn^{2+} + 106\ CO_2 + 8\ N_2 + 366\ H_2O + H_3PO_4$

-3090 kJ/mol

Oxidation by Nitrate

$(CH_2O)_{106}(NH_3)_{16}(H_3PO_4) + 84.4\ HNO_3 \rightarrow$
$106\ CO_2 + 42.2\ N_2 + 16\ NH_3 + 148.4\ H_2O + H_3PO_4$

-2750 kJ/mol

Oxidation by Iron-Oxides

$(CH_2O)_{106}(NH_3)_{16}(H_3PO_4) + 212\ Fe_2O_3 + 848\ H^+ \rightarrow$
$424\ Fe^{2+} + 106\ CO_2 + 16\ NH_3 + 530\ H_2O + H_3PO_4$

-1410 kJ/mol

anoxic

Oxidation by Sulfate

$(CH_2O)_{106}(NH_3)_{16}(H_3PO_4) + 53\ SO_4^{2-} \rightarrow$
$106\ CO_2 + 16\ NH_3 + 53\ S^{2-} + 106\ H_2O + H_3PO_4$

-380 kJ/mol

Methane Fermentation

$(CH_2O)_{106}(NH_3)_{16}(H_3PO_4) \rightarrow$
$53\ CO_2 + 53\ CH_4 + 16\ NH_3 + H_3PO_4$

-350 kJ/mol

Energetically

most favorable

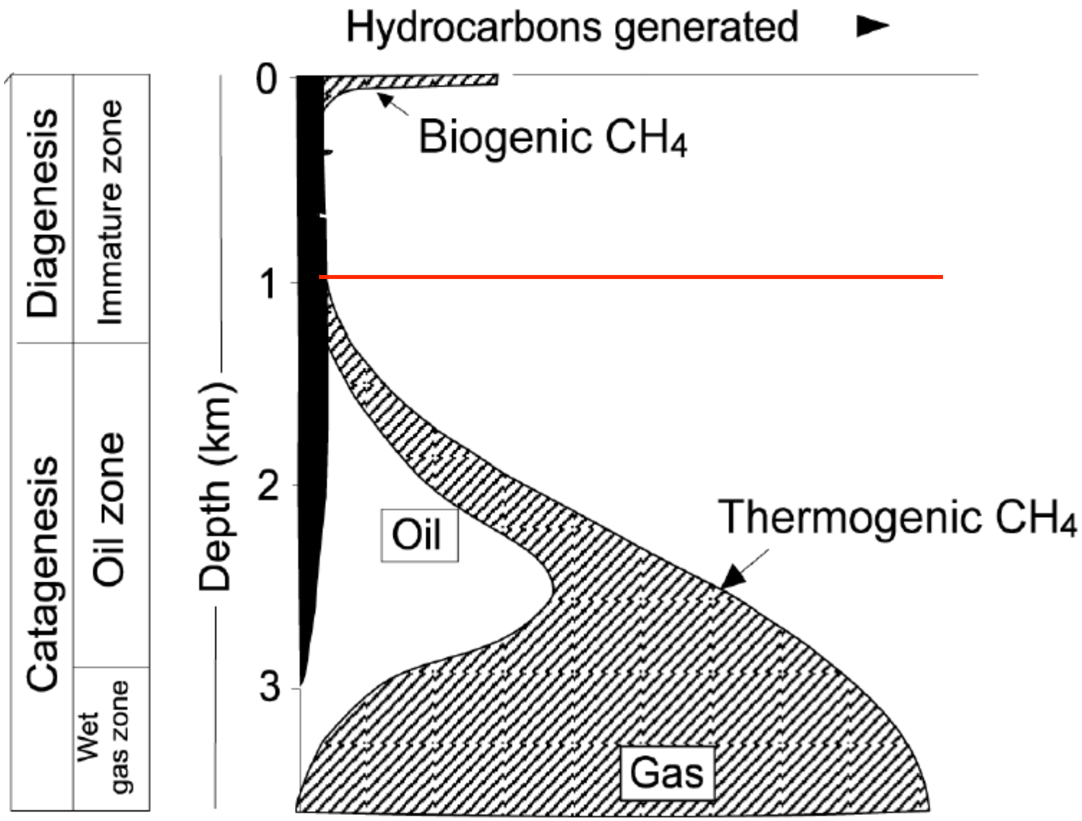
least favorable

Hydrocarbon Generation

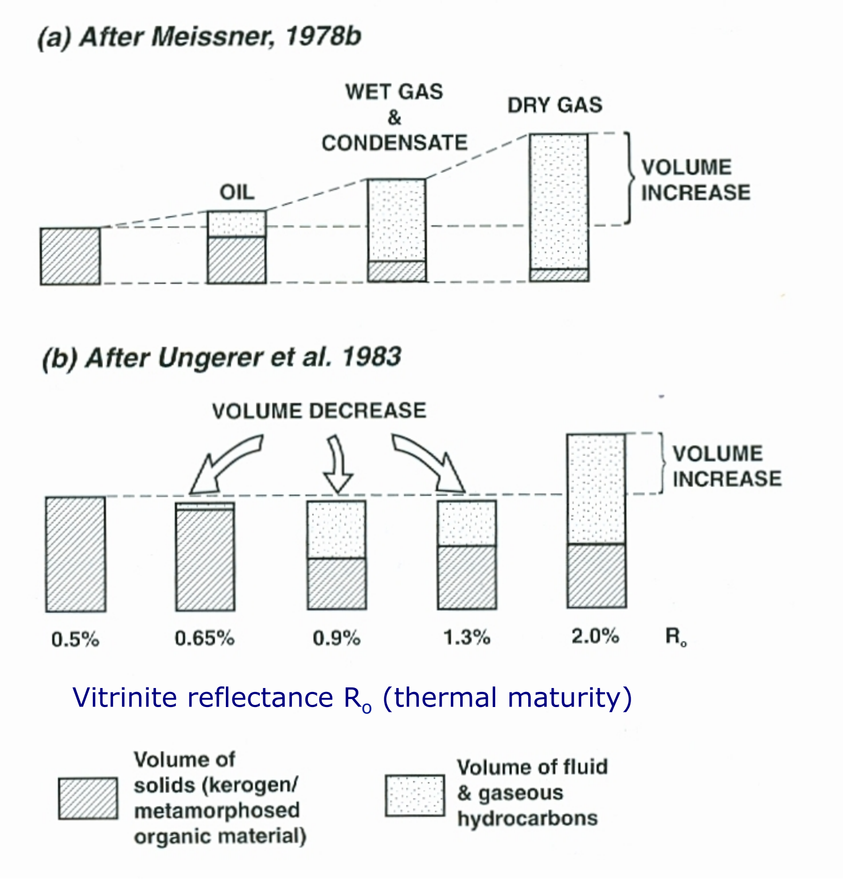
• Thermogenic Methane

Deeper in the sediment, **thermal alteration** of organic matter generates methane and higher order hydrocarbons by **catagenesis**.

Catagenesis occurs within the temperature range of 50° to 200° C, and gases (methane to butane) are produced at rates that are proportional to temperature.

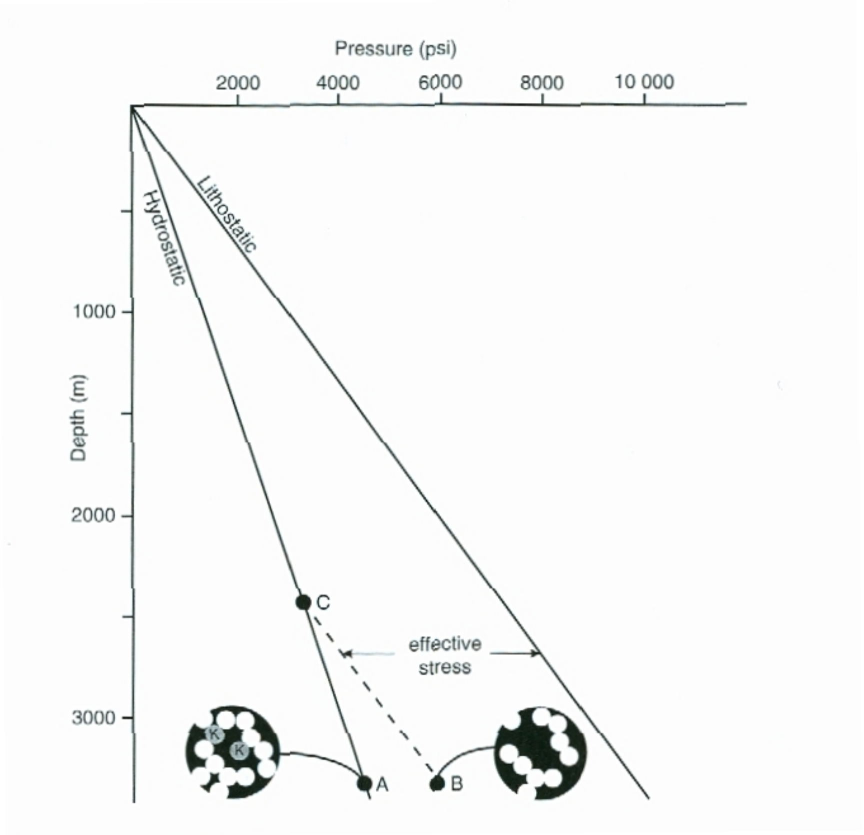


Tissot and Welte, Petroleum Formation and Occurrence, Springer-Verlag, 1992.



Vitrinite reflectance R_o (thermal maturity)

Estimation of volume change when Type II Kerogene matures to produce oil and gas.



Estimation of the overpressure created by disequilibrium compaction in a maturing source rock. Half of the initial Kerogene (10% by volume) is transformed into liquid products, thereby increasing the porosity from 13 to 18%. The effect is to transfer the part of the overburden supported by the original kerogene onto the pore fluid.

Swarbrick and Osborne, 1998

Factors that determine excess pore water pressure and influence the consolidation state

The most common factors are:

- **Stress-related factors**
    - *Rapid sedimentation rate*
    - *Tectonic stress*
- **Fluid volume increase mechanisms**
    - *Temperature increase*
    - *Mineral transformation*
    - *Hydrocarbon Generation*
- **Fluid movement mechanisms**
    - Osmosis
    - Hydraulic head
    - Hydrocarbon buoyancy

(**Swarbrick and Osborne, 1998**. Bryant et al., 1974; Sangrey, 1977; Arthur et al., 1980; Demaison & Moore, 1980; Bryant et al.,1981).

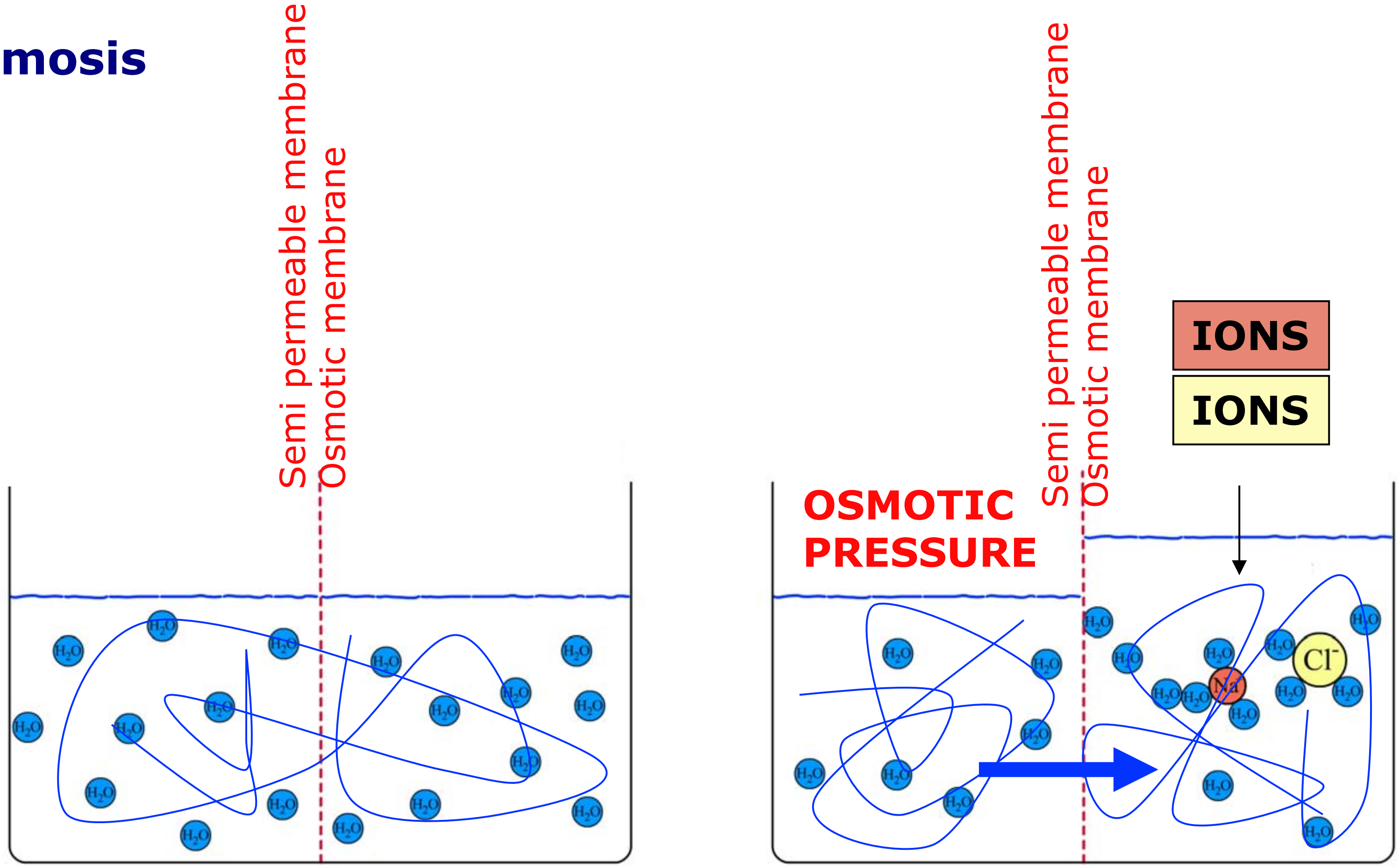
Swarbrick, and Osborne, 1998

Osmosis

Osmosis is the diffusion of a solvent (e.g. water) through a membrane (osmotic membrane, or semi-permeable membrane) from a solution of low solute concentration to a solution with high solute concentration, up a solute concentration gradient.

A solvent moves without input of energy, across a semi permeable membrane (permeable to the solvent, but not the solute) separating two solutions of different concentrations.

Osmosis


Semi permeable membrane
Osmotic membrane
H_2O
Semi permeable membrane
Osmotic membrane
OSMOTIC PRESSURE
IONS
IONS
Na^+
Cl^-

Osmosis

As a consequence of pore water expulsion due to consolidation, pore water salinity often increases with subsurface depth in a sedimentary basin due to

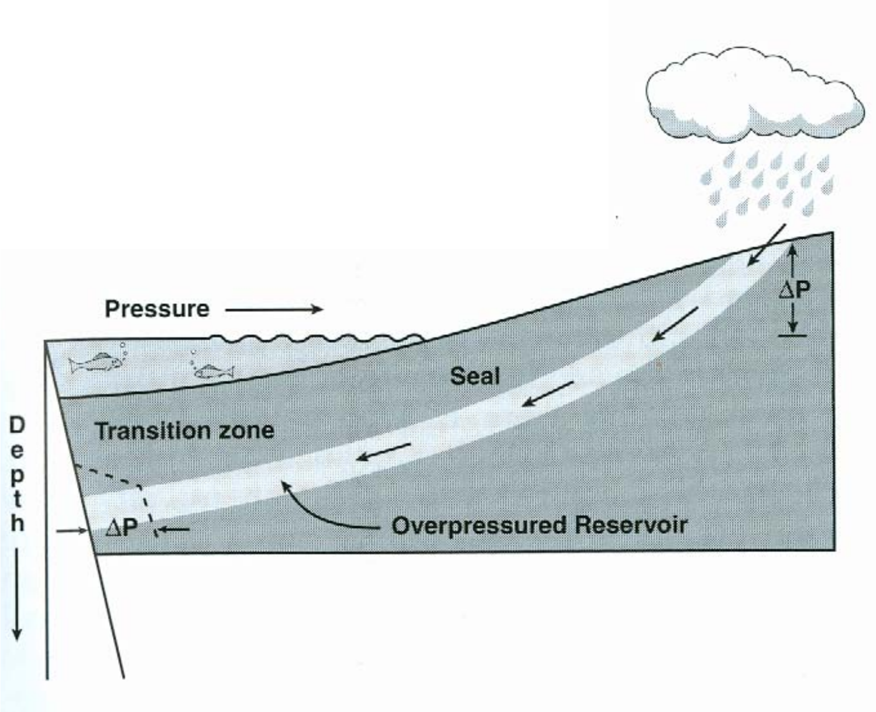
REVERSE OSMOSIS.

Reverse osmosis occurs when a pressure head forces brines through an osmotic membrane, allowing the solvent to pass and retaining the solute.

Reverse osmosis is therefore considered as a possible sink of Na^+ and Cl^+ ions from seawater.

Graf, 1982

Hydraulic head



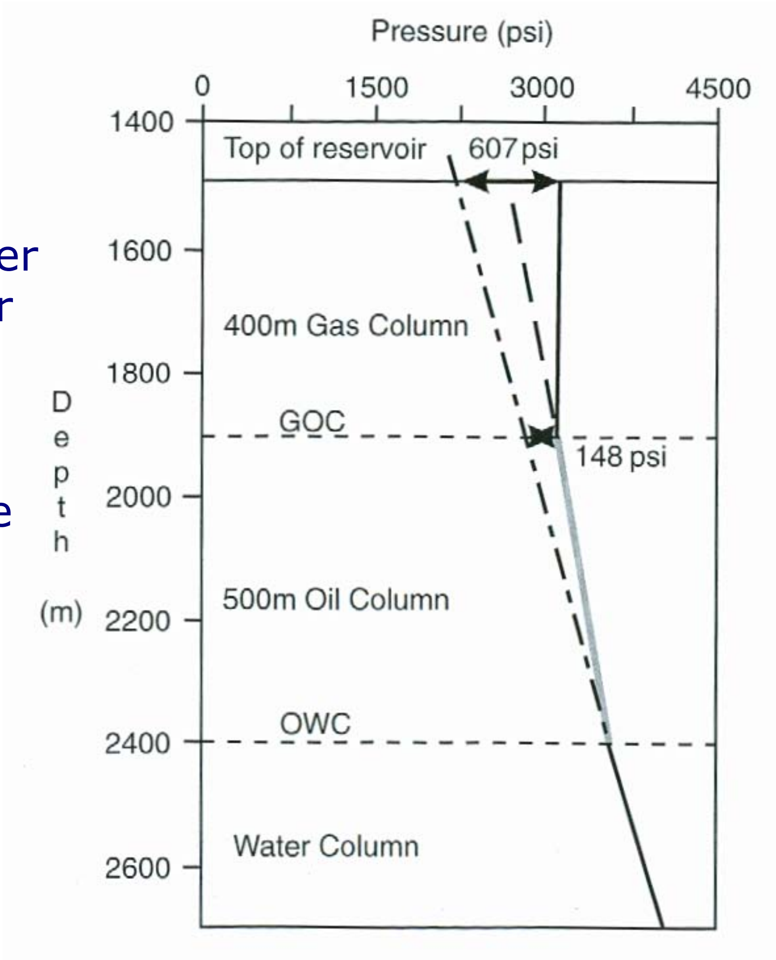
Swarbrick, and Osborne, 1998

Hydrocarbon buoyancy

- All gasses and oils have lower density than water
- They have a lower pressure gradient than water
- Because overpressure is an excess pressure above hydrostatic for a given depth,

There is always a certain amount of overpressure ina a reservoir.

This is a **local** overpressure



Swarbrick, and Osborne, 1998

APPENDIX. PRESSURE UNITS CONVERSION

The SI unit for **pressure** is the Pascal (Pa)

1 Pa = one Newton per square metre ($N \cdot m^{-2}$ or $kg \cdot m^{-1} \cdot s^{-2}$).

Non-SI measures such as *pound* per square inch (**psi**) and ***bar***

$$1\ \text{Pa} = 10^{-5}\ \text{bar} = 145.04 \cdot 10^{-6}\ \text{psi}$$

$$1\ \text{bar} = 100{,}000\ \text{Pa} = 14.504\ \text{psi}$$

$$1\ \text{psi} = 6{,}894.76\ \text{Pa} = 68.948 \cdot 10^{-3}\ \text{Bar}$$

$$1\ \text{bar} = 0.98692\ \text{atmospheres}$$

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