Geodynamics of sedimentary basins
Magdalena Scheck-Wenderoth
magdalena.scheck@gfz-potsdam.de

who am I?
Magdalena Scheck-Wenderoth
Professor for Basin Analysis at RWTH Aachen
in joint appointment with
German Research Centre for Geosciences GFZ
Helmholtz Centre Potsdam
Director Department 4: Geosystems
Head Sect. 4.5: Basin Modelling
Section 4.5 Basin Modelling

geodynamic processes in sedimentary basins: quantify mass and energy fluxes

integrate observations into 3D subsurface models for key areas from regional to reservoir scale

4D process simulation of Thermal-Hydraulic-Mechanical processes across scales
observation-based models

observations

- Field
- wells
  - Stratigraphy
  - Lithology
  - Temperature
  - Paleo indicators
- Seismic
- Potential fields
- Seismology

Interpretation & Structural modelling

- Structural analysis
- Potential field modelling
- THM modelling
- Subsidence analysis
- Deformation restoration

Literature –(text)books


K. Bjørlykke (ed.), *Petroleum Geoscience: From Sedimentary Environments to Rock Physics*, 1, DOI 10.1007/978-3-642-02332-3_1, © Springer-Verlag Berlin Heidelberg 2010


http://www.geolsoc.org.uk/ks3/gsi/education/resources/rockcycle.html

Xie, X. and Heller, P. L. Plate tectonics and basin subsidence history. GSA Bulletin; January/February 2009; v. 121; no. 1/2; p. 55–64; doi: 10.1130/B26398.1

Literature –+relevant articles theme-specific
Overview

- Short Intro to Basin types
- Mechanisms of basin formation
- Heat transport in sedimentary basins
- Geophysical characteristics of
  - rift basins:
  - passive margins
  - intracontinental basins:
  - foreland basins:
  - strike-slip basins

Sedimentary basins

- are the subsiding areas where sediments accumulate to form stratigraphic successions=> ARCHIVES
- form in response to different mechanisms in different plate-tectonic settings
- have characteristic geophysical fingerprint, subsidence and temperature histories
- represent low pressure-high temperature reactors in which accumulated sediments experience changing conditions
- host georesources (water, petroleum, heat)
Sedimentary basins why should we care?
global thickness distribution of sediments

Sediment Thickness (km) at 1x1°

Gabi Laske and Guy Masters
https://igppweb.ucsd.edu/~gabi/sediment.html

some sedimentary basins

Atlantic
Wannsee
valley Himalaya
Lake Baikal
a few questions...

which sedimentary basins do you know?

which types of sedimentary basins do you know?

rift basin,
intracontinental basin,
passive continental margins,
foreland basin,
subduction trench,
fore-arc- basin,
back-arc- basin,
pull-apart- basin,
piggy-back - basin
Basin Types

many classifications...

after tectonic setting:
  divergent => extensional basins
  convergent => collisional basins
  strike-slip => transtensional basins

after formation mechanism
  extensional basins
  flexural basins

after lithosphere substrate
  oceanic
  continental

after petroleum prospectivity

Where do basins form?
basins in their plate-tectonic setting

Wilson cycle
Basin Forming Mechanisms?
in the end...it's almost all about densities & heat flow ... (or their changes)

Some basics before looking at basins...

- Isostasy = f(load equilibria) = f(density) => controls stresses induced by loads and therefore subsidence and related effects (changes in physical properties)
- Temperature => controls physical properties and rheology of rocks
compositional/rheological boundaries in the earth

Crust = sediments + upper silicic cryst. crust + lower mafic cryst. crust

upper mantle: peridotite

The thermal lithosphere-asthenosphere boundary (LAB)

Fig. 2.17: The geotherm for the lithosphere and the solidus for peridotite. The geotherm approaches the solidus at the base of the lithosphere, where partial melting may take place. The geothermal gradient is calculated using 1D conduction with a thermal conductivity of 3.0 W/m·K, a basal heat flow of 80 mW/m², and a surface temperature of 25°C. Pressure is calculated using a geostatic 40 km thick crust with density 2750 kg/m³ and a mantle lithosphere extending to a depth of 200 km with density 3300 kg/m³. The geotherm is adiabatic in the asthenosphere and takes an 0.2°C/m depth. The solidus for peridotite is given by T = 1500 + 0.12p where pressure (p) is in kbar. Note that the geotherm approaches the solidus near the lithosphere-asthenosphere boundary, leading to partial melting.
how to make a basin?

The ideal lithosphere in perfect equilibrium with the underlying asthenosphere

Almost nowhere on earth this is the case because the lithosphere is laterally and vertically heterogeneous + under stress (extension/compression/buoyancy) and deforms.

LAB: ~1300°C

\[ \rho_{\text{crust}} = 2800 \text{ kg/m}^3 \]
\[ m_{\text{crust}} = 30000 \]
\[ \rho_{\text{th}} = 3300 \text{ kg/m}^3 \]
\[ m_{\text{th}} = 70000 \]
\[ \rho_{\text{asth}} = 3200 \text{ kg/m}^3 \]
how to make a basin?

- stretch/extend
- bend/flex

mechanisms of basins formation

extension of lithospheric plates

flexure due to vertical or horizontal loads

\[ \rho_{\text{crust}} = 2800 \text{ kg/m}^3 \]
\[ m_{\text{crust}} = 30000 \]
\[ \rho_{\text{asth}} = 3200 \text{ kg/m}^3 \]
\[ m_{\text{asth}} = 70000 \]
\[ \rho_{\text{lith}} = 3300 \text{ kg/m}^3 \]
\[ m_{\text{lith}} = 70000 \]
mechanisms of basins formation

extension of lithospheric plates

• active vs passive rift models
• pure shear vs simple shear models
• uniform vs depth-dependent stretching

flexure due to vertical or horizontal loads

• compressive lithosphere folding (horiz. compression)
• orogenic loading=foreland basins
• basal loading (phase transitions; subducted slab remains; down-welling in the deeper mantle)

Isostasy

is the concept that the elevation of the Earth's surface seeks a balance between the weight of lithospheric rocks and the buoyancy of asthenospheric "fluid" (nearly-molten rock).

Archimedes’ Principle

Simple example: iceberg
**Isostasy** Lithosphere with lateral load variations

Local Isostasy

\[ \rho_{\text{lith}} y_{\text{lith}} g = \rho_{\text{asth}} y_{\text{asth}} \]

\[ \rho_{\text{lith}} < \rho_{\text{asth}} \]

Lithosphere

Depth of isostatic compensation

---

**Isostasy** Lithosphere with lateral load variations

Local Isostasy

\[ \rho_{\text{lith}} y_{\text{lith}} = \rho_{\text{asth}} y_{\text{asth}} \]

\[ \rho_{\text{crust}} y_{\text{crust}} + \rho_{\text{lithm}} y_{\text{lithm}} = \rho_{\text{asth}} y_{\text{asth}} \]

Lithosphere

Depth of isostatic compensation
Isostasy
Lithosphere with lateral load variations

Local isostasy: 1855

Pratt
No crustal root below orogens

Isostasy is achieved by lateral density variations
\[ p_1 < p_3 < p_2 < p_4 \]

Depth of isostatic compensation

G. H. Pratt hypothesized that the density of the crust varies, allowing the base of the crust to be the same everywhere. Sections of crust with high mountains, therefore, would be less dense than sections of crust where there are lowlands. This applies to instances where density varies, such as the difference between continental and oceanic crust.

George Bedell Airy proposed that the density of the crust is everywhere the same and the thickness of crustal material varies. Higher mountains are compensated by deeper roots. This explains the high elevations of most major mountain chains, such as the Himalayas.

Airy
Crustal root below orogens

Isostasy is achieved by lateral thickness variations
\[ y_1 > y_3 > y_2 > y_4 \]

G. H. Pratt hypothesized that the density of the crust varies, allowing the base of the crust to be the same everywhere. Sections of crust with high mountains, therefore, would be less dense than sections of crust where there are lowlands. This applies to instances where density varies, such as the difference between continental and oceanic crust.

Isostasy is achieved by lateral thickness variations
\[ p_1 = p_3 = p_2 = p_4 \]

mantle
Isostasy

Lithosphere with lateral load variations

Local isostasy

Isostasy is achieved by lateral thickness variations
\( \rho_1 = \rho_3 = \rho_2 = \rho_4 \)
\( y_1 > y_3 > y_2 > y_4 \)

Airy
Crustal root below orogens

Pratt
No crustal root below orogens

\( \rho_{\text{lith}} y_{\text{lith}} = \rho_{\text{asth}} y_{\text{asth}} \)

Depth of isostatic compensation

Flexural isostasy

Lithosphere has a finite strength and acts like an elastic beam
Bends down if loaded
Wavelegnth and amplitude of deflection is a function of rigidity
**Extension of lithospheric plates**

![Diagram of lithospheric plates](image)

**Initial Lithosphere Structure**

\[ d_{\text{initial}} \times l_{\text{initial}} = d_{\text{stretched}} \times l_{\text{stretched}} \]

\[ \beta = \frac{d_{\text{initial}}}{d_{\text{stretched}}} = \frac{l_{\text{stretched}}}{l_{\text{initial}}} \]

**Stretching factor (\( \beta \))**

- is the ratio of the initial lithosphere thickness to the final (stretched) thickness of the lithosphere.

- example: \( \beta = 2 \) means, that lithosphere has been thinned to
  
  (a) \( \frac{1}{4} \)
  
  (b) \( \frac{1}{2} \)
  
  (c) none of the two?
stretching factor ( $\beta$ )

- is the ratio of the initial lithosphere thickness to the final (stretched) thickness of the lithosphere.

- example: $\beta=2$ means, that lithosphere has been thinned to

  (a) $\frac{1}{4}$
  (b) $\frac{1}{2}$
  (c) none of the two?

Isostatic consequences of lithosphere extension

If plate boundary forces are absent, at the Lithosphere-Asthenosphere-Boundary, the load $P_{LAB}$ in isostatic equilibrium is

$$P_{LAB} = \rho_c y_c^* g + \rho_m y_m^* g$$

Load of crust

Load of lithospheric mantle

Level of isostatic compensation
Isostatic consequences of lithosphere extension

after stretching by $\beta$ the load $P_{LABS}$ is reduced

$$P_{LABS} = \rho_c \beta^{-1} y_c g + \rho_m \beta^{-1} y_m g$$

Isostatic compensation is accomplished by rising asthenosphere $y_a$ with $\rho_a$

level of isostatic compensation

for isostatic equilibrium:

$$\rho_c y_c g + \rho_m y_m g = \rho_c \beta^{-1} y_c g + \rho_m \beta^{-1} y_m g + \rho_a y_a g$$

$$\rho_c y_c + \rho_m y_m = \rho_c \beta^{-1} y_c + \rho_m \beta^{-1} y_m + \rho_a y_a$$
Isostatic consequences of lithosphere extension

How much Asthenosphere has to flow in, if a lithosphere of initially 100 km thickness with a crust of 30 km is stretched by $\beta=2$?

\[ \rho_c^*y_c + \rho_m^*y_m = \rho_c^*\beta^{-1}y_c + \rho_m^*\beta^{-1}y_m + \rho_a^*y_a \]

\[ (\rho_c^*y_c + \rho_m^*y_m - \rho_c^*\beta^{-1}y_c - \rho_m^*\beta^{-1}y_m)/\rho_a = y_a \]
Isostatic consequences of lithosphere extension

How much Asthenosphere has to flow in, if a lithosphere of initially 100 km thickness with a crust of 30 km is streched by $\beta=2$?

$$\rho_c^*y_c + \rho_m^*y_m = \rho_c^*\beta^{-1}y_c + \rho_m^*\beta^{-1}y_m + \rho_a^*y_a$$

$$\frac{(\rho_c^*y_c + \rho_m^*y_m - \rho_c^*\beta^{-1}y_c - \rho_m^*\beta^{-1}y_m)}{\rho_a} = y_a$$

$d_c=30\text{km}$
$d_m=70\text{km}$
$\beta^{-1}d_c = 30/2=15\text{km}$
$\beta^{-1}d_m = 70/2=35\text{km}$

$\rho_c=2800\text{kg/m}^3$
$\rho_m=3300\text{kg/m}^3$
$\rho_a=3200\text{kg/m}^3$

$$(2800\text{kg/m}^3\times30\text{km} + 3300\text{kg/m}^3\times70\text{km} - 2800\text{kg/m}^3\times15\text{km} - 3300\text{kg/m}^3\times35\text{km}) 3300\text{kg/m}^3 = y_a$$

$49218,75\text{km} = y_a$
Isostatic consequences of lithosphere extension

what did we forget????

For a water-filled basin:

$$\rho_c y_c + \rho_m y_m = \rho_c \beta^{-1} y_c + \rho_m \beta^{-1} y_m + \rho_a y_a + \rho_w y_w$$

For a sediment-filled basin:

$$\rho_c y_c + \rho_m y_m = \rho_c \beta^{-1} y_c + \rho_m \beta^{-1} y_m + \rho_a y_a + \rho_s y_s$$

For a basin filled with water and sediment:

$$\rho_c y_c + \rho_m y_m = \rho_c \beta^{-1} y_c + \rho_m \beta^{-1} y_m + \rho_a y_a + \rho_s y_s + \rho_w y_w$$

The basin fill!!!!!
**Isostatic consequences of lithosphere extension**

For $\beta = 2$ => 30/2 = 15 km of the crust are filled with water

$$\rho_c \cdot y_c + \rho_m \cdot y_m = \rho_c \cdot \beta^{-1} y_c + \rho_m \cdot \beta^{-1} y_m + \rho_s \cdot y_s + \rho_w \cdot y_w$$

$$(\rho_c \cdot y_c + \rho_m \cdot y_m - \rho_s \cdot y_s - \rho_w \cdot y_w) / \rho_c \cdot \beta = y_a$$

$$(2800 \text{ kg/m}^3 \cdot 30 \text{ km} + 3300 \text{ kg/m}^3 \cdot 70 \text{ km} - 1000 \text{ kg/m}^3 \cdot 15 \text{ km} - 2800 \text{ kg/m}^3 \cdot 15 \text{ km} - 3300 \text{ kg/m}^3 \cdot 35 \text{ km}) / 3300 \text{ kg/m}^3 = y_a$$

44531,25 km = $y_a$

---

**Isostatic consequences of lithosphere extension**

If the air-filled basin receives 5 km Sediment with $\rho_{sed} = 2400 \text{ kg/m}^3$

$$42343,75 = y_a$$

......

water-filled 44531,25 km = $y_a$

air-filled 49218,75 km = $y_a$
### Isostatic consequences of some processes...

<table>
<thead>
<tr>
<th>$\rho_c$</th>
<th>$\rho_m$</th>
<th>$\rho_a$</th>
<th>$\rho_w$</th>
</tr>
</thead>
<tbody>
<tr>
<td>2800 kg/m³</td>
<td>3300 kg/m³</td>
<td>3200 kg/m³</td>
<td>1000 kg/m³</td>
</tr>
</tbody>
</table>

How much isostatic rebound would a 200m sea level drop cause?

Load that needs to be compensated by athenosphere:

\[ \rho_W \times y_W = \rho_A y_A \]

\[ \rho_W \times y_W / \rho_A = y_A \]

### Isostatic consequences of sea level drop

<table>
<thead>
<tr>
<th>$\rho_c$</th>
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<th>$\rho_a$</th>
<th>$\rho_w$</th>
</tr>
</thead>
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<tr>
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<td>3300 kg/m³</td>
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<td>1000 kg/m³</td>
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</tbody>
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How much isostatic rebound would a 200m sea level drop cause?

Load that needs to be compensated by athenosphere:

\[ \rho_W \times y_W = \rho_A y_A \]

\[ \rho_W \times y_W / \rho_A = y_A \]

\[ 1000kg/m^3 \times 200m / 3200kg/m^3 = 62.5m \]
Isostatic consequences of sea level drop

How much isostatic rebound would a 200m sea level drop cause? 62.5 m

How much isostatic rebound would the melting of a 2km thick ice sheet cause?

\[ \rho_W \cdot y_W / \rho_a = y_A \]
Isostatic consequences of ice melting

\[ \rho_{c} = 2800 \text{ kg/m}^3 \]
\[ \rho_{m} = 3300 \text{ kg/m}^3 \]
\[ \rho_{a} = 3200 \text{ kg/m}^3 \]
\[ \rho_{w} = 1000 \text{ kg/m}^3 \approx \frac{1}{3} \rho_{a} \]

How much isostatic rebound would a 200m sea level drop cause?

How much isostatic rebound would the melting of a 2km thick ice sheet cause?

\[ \rho_{w} \cdot y_{W} / \rho_{A} = y_{A} \]
\[ 1000 \text{ kg/m}^3 \times \frac{2000}{3200} = 625 \text{ m} \]

Can 2000m of erosion/denudation be explained by sea level changes or deglaciation?
Isostatic consequences of sea level and glaciations

\[ \rho_c = 2800 \text{ kg/m}^3 \]
\[ \rho_m = 3300 \text{ kg/m}^3 \]
\[ \rho_s = 3200 \text{ kg/m}^3 \]
\[ \rho_w = 1000 \text{ kg/m}^3 \]
\[ \approx \frac{1}{3} \rho_a \]
\[ \rho_{sed} = 2300 \text{ kg/m}^3 \]
\[ \approx \frac{2}{3} \rho_a \]

How much isostatic rebound would a 200m sea level drop cause? \(~60\) m

How much isostatic rebound would the melting of a 2km thick ice sheet cause? \(~600\) m

Can 2000m of denudation be explained by sea level changes or deglaciation? \textbf{NO!}

Isostatic consequences of erosion

\[ \rho_c = 2800 \text{ kg/m}^3 \]
\[ \rho_m = 3300 \text{ kg/m}^3 \]
\[ \rho_s = 3200 \text{ kg/m}^3 \]
\[ \rho_w = 1000 \text{ kg/m}^3 \]
\[ \approx \frac{1}{3} \rho_a \]
\[ \rho_{sed} = 2300 \text{ kg/m}^3 \]
\[ \approx \frac{2}{3} \rho_a \]

How much isostatic rebound would the erosion of 1 km of sediments cause?

\[ \rho_{sed} \cdot \frac{y_{sed}}{\rho_A} = y_A \]
\[ 2300 \text{ kg/m}^3 \times \frac{1000 \text{ m}}{3200 \text{ kg/m}^3} = 718 \text{ m} \]
Isostatic consequences of erosion

\[ \rho_c = 2800 \text{kg/m}^3 \]
\[ \rho_a = 3200 \text{kg/m}^3 \]
\[ \rho_s = 1000 \text{kg/m}^3 \]
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How much isostatic rebound would the erosion of 1 km of sediments cause?

\[ \rho_{sed} \cdot y_{sed} / \rho_a = y_A \]

\[ 2300 \text{kg/m}^3 \times 1000 \text{m} / 3200 \text{kg/m}^3 = 718 \text{ m} \]

But! Isostasy may be just one part of a sequence of coupled processes.

Fig. 9.14 Schematic illustration of the effects of initial water depth on sediment thickness. (i) Sedimentary basin with an initial water depth of 1 km (ii) undergoes 1 km of tectonic subsidence (iii) and is filled with sediment to a depth of 13 km (iv). (v) Sedimentary basin initially at sea level (vi) undergoes 1 km of tectonic subsidence (vii) and accumulates c.3 km of sediment (viii).
Rules to remember about isostasy:

- Ocean crust is thin and dense (5-10 km, density of ~3000 kg/m³).
- Continental crust is thick and less dense (30-50 km density ~ 2800 kg/m³).
- Less dense crust rises higher (continental crust, mountains) than does more dense crust (ocean crust). Oceans are low and continents are high.
- Loading the crust causes it to sink (sediments or ice sheet on top).
- Unloading the crust causes it to rise (as in eroding a mountain down).

Isostatic consequences of lithosphere extension

Isostacy is a static concept.

Geology of basins takes time.

-> How to integrate time dependence?

-> McKenzie:
changes in temperature (and the related effects on density) over time are the key!
Heat Transfer

Heat flow or transfer is a process achieved in the lithosphere dominantly by conduction.

The thermal lithosphere-asthenosphere boundary (LAB)

Fig. 2.17  The geotherms for the lithosphere and the solidus for peridotite. The geotherm approaches the solidus at the base of the lithosphere, where partial melting may take place. The geothermal gradient is calculated using 1D conduction with a thermal conductivity of 3.54 W m⁻¹ K⁻¹, a local heat flow of 40 mW m⁻², and a surface temperature of 0°C. Pressure is calculated using a geostatic 40 km thick crust with density 2.75 kg m⁻³ and a mantle lithosphere extending to a depth of 250 km with density 3300 kg m⁻³. The geotherm is adiabatic in the asthenosphere and taken as 0.5°C km⁻¹. The solidus for peridotite is given by T = 1500 + 0.12p where pressure (p) is in MPa. Note that the geotherm approaches the solidus near the lithosphere-asthenosphere boundary, leading to partial melting.
heat in basins...

controlled by

heat sources

conduction

convection

heat transport mechanisms

LAB ~1300°C

• heat input from asthenosphere

~10°C

gradT
heat sources

- heat input from asthenosphere
- radiogenic heat produced in the lithosphere

LAB ~1300°C

crust

• contributes significantly to heat generation in the crust
• most rocks contain radioactive minerals
• concentrations are low, but due to large thickness may affect the temperature distribution (source or sink in the heat equation)
• uranium, potassium, thorium

\[ A = \rho \left( 3.35 c_{\text{ck}} + 9.79 c_{\text{cu}} + 2.64 c_{\text{crh}} \right) \times 10^{-5} \]

(Schmucker, 1969)

- heat production
- \( \rho \) density
- \( c \) heat generation constants for known concentrations

low \( \rightarrow \) carbonates & carbonates
medium \( \rightarrow \) sandstones
high \( \rightarrow \) granite, silt- and claystones
heat producing elements in sedimentary rocks

<table>
<thead>
<tr>
<th>Rock type</th>
<th>U (ppm)</th>
<th>Th (ppm)</th>
<th>K (%)</th>
<th>Th/U</th>
<th>Density (g kg(^{-1}))</th>
<th>Heat generation (\mu W/m(^2))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Carbonates</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Limestone</td>
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<td>1.5</td>
<td>0.3</td>
<td>0.75</td>
<td></td>
<td>0.62</td>
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<tr>
<td>Dolomite</td>
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<td>0.8</td>
<td>0.7</td>
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<tr>
<td>Evaporites</td>
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<tr>
<td>Salt</td>
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<td>0.01</td>
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<tr>
<td>Quartzite</td>
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<td></td>
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<td>Graywacke</td>
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<td>1.3</td>
<td>3.5</td>
<td></td>
<td>0.99</td>
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<tr>
<td>Deep sea sediments</td>
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<td>11.0</td>
<td>2.5</td>
<td>5.2</td>
<td></td>
<td>1.3</td>
</tr>
</tbody>
</table>

heat transport mechanism?

- conduction
- advection
- convection
- by diffusion
- pressure driven
- bouyancy-driven

relevant at different scales
Heat Transfer by Conduction

Diffuse and therefore SLOW process by which the kinetic energy is transferred by interatomic or intermolecular collision

Conductive heat flow

Fourier’s law

\[ Q = -K \frac{dT}{dy} \]
\[ Q = -K \cdot \text{grad}T \]

Q: heat flow
K: thermal conductivity
T: temperature
y: coordinate in the direction of the temperature variation (e.g. depth)

Heat flow is in units of \( \text{mW} \ \text{m}^{-2} \) or \( \text{cal cm}^{-2} \ \text{s}^{-1} \)
HFU (heat flow units) = \( 10^{-6} \) cal cm\(^2\) s\(^{-1}\) or 41.84 mW m\(^{-2}\)
Conductive Heat Transfer

\[
\frac{\partial}{\partial x} \left( K_x \frac{\partial T}{\partial x} \right) + \frac{\partial}{\partial y} \left( K_y \frac{\partial T}{\partial y} \right) + \frac{\partial}{\partial z} \left( K_z \frac{\partial T}{\partial z} \right) = \rho c \frac{\partial T}{\partial t} + A
\]

K – thermal conductivities
\(c\) – specific heat
\(\rho\) – density
S – source or sink term (radiogenic heat, convection)

Any loss or gain in heat in a unit volume is balanced by a temperature change multiplied with volumetric heat capacity (Gretener, 1981).

radiogenic heat enters as a source term A

If conduction has time enough to equilibrate on the scale of the lithosphere (\(dT/dt=0\))

\[
\Rightarrow \text{steady state conditions:}
\]

\[
\frac{\partial}{\partial x} \left( K_x \frac{\partial T}{\partial x} \right) + \frac{\partial}{\partial y} \left( K_y \frac{\partial T}{\partial y} \right) + \frac{\partial}{\partial z} \left( K_z \frac{\partial T}{\partial z} \right) = \rho c \frac{\partial T}{\partial t} + A
\]

A – radiogenic heat production [mW/m³]
T – temperature [°C]
k- conductivity [W/mK]
Heat Transfer by Conduction

**Thermal Conductivity of rocks**

= change in temperature per unit length for a specific amount of heat

High thermal conductivity => low thermal gradient
Low thermal conductivity => high thermal gradient

**thermal conductivity of sedimentary rocks**

Thermal conductivity \( \lambda \) [W/(m·K)−1]

heat \( W \) transported over a distance [m] with a drop in Temperature [°C or K]

Bulk thermal conductivity \( \lambda_b = \phi \lambda_f + (1-\phi)\lambda_S \)

- Porosity and fluid content (water, brine, oil, gas)
- Function of depth (porosity loss with burial - compaction)
- Mineralogy of the framework grains
- Type and amount of material in the matrix (clay minerals)
thermal conductivity of sedimentary rocks

<table>
<thead>
<tr>
<th>Rock type</th>
<th>( \lambda ) [( \text{W} \cdot \text{m}^{-1} \cdot \text{K}^{-1} )]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water (pure fluid)</td>
<td>0.6</td>
</tr>
<tr>
<td>Quartz</td>
<td>7.7</td>
</tr>
<tr>
<td>Calcite</td>
<td>3.3</td>
</tr>
<tr>
<td>Amphibole</td>
<td>6.3</td>
</tr>
<tr>
<td>Slate</td>
<td>1.2 - 3.6</td>
</tr>
<tr>
<td>Sandstone</td>
<td>1.5 - 4.2</td>
</tr>
<tr>
<td>Limestones</td>
<td>2.0 - 3.4</td>
</tr>
<tr>
<td>Basalt</td>
<td>1.3 - 2.9</td>
</tr>
<tr>
<td>Granite</td>
<td>2.4 - 3.8</td>
</tr>
<tr>
<td>Peridotite</td>
<td>3.0 - 4.5</td>
</tr>
</tbody>
</table>

Thermal conductivity of some common rocks (after Brögger & Yuen, 1989)

\[ \text{high } \lambda \Rightarrow \text{efficient heat transfer} \]
\[ \Rightarrow \text{low thermal gradient} \]

\[ \text{low } \lambda \Rightarrow \text{reduced heat transfer} \]
\[ \Rightarrow \text{high thermal gradient} \]
Fig. 10.2: The influence of internal heat generation per unit volume in the sedimentary column $q$ and thermal conductivity $k$ on the distribution of temperature with depth $T(z)$. The different curves were calculated for $k_0$ values for a thickness of the heat-producing zone of $1000 \text{ m}^2$, a basal heat flux $q_b$ of $0.01 \text{ kW/m}^2$ and a surface temperature $T_0$ of $10^\circ \text{C}$. Reproduced with permission from Eidedix Technix.

blanketing effects
relevant properties

- **sediments:** mostly insulating, moderate radiogenic heat production, pore fluids
  - ⇒ heat transport,
  - ⇒ „thermal blanketing“

- **crust:** conductive, high radiogenic heat production
  - ⇒ heat transport,
  - ⇒ heat budget

- **lithospheric mantle:** conductive, low radiogenic heat production
  - ⇒ heat transport

- **depth to LAB:** thermal gradient,
  - ⇒ heat budget

---

**typical HF values**

Allen & Allen, 2013, Fig. 3.8

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Allen & Allen, 2013, Fig. 10.31
Observations

- Sediment fill can be a few 100m and up to >22km thick

- Structure of sediments often indicates 2 major phases:
  - Syn-rift: normal faults and grabens define sub-basins => initial, **FAST** subsidence
  - Post-rift: discordantly on synrift, onlaps on sub-basins, spatial widening of subsiding area, continuous layering, **decreasing rate of subsidence with time** => thermal subsidence

- **Crust**: thinned, seismic velocities and densities of the crust indicate changes of crustal configuration and composition in response to extension, strongly variable for different basins

- **Lithospheric mantle**: strongly variable for different basins
Isostatic consequences of lithosphere extension

Isostacy is a static concept.

Geology of basins takes time.

-> How to integrate time dependence?

-> McKenzie: changes in temperature (and the related effects on density) over time are the key!

Extensional rift models...

Uniform and symmetric lithospheric thinning by $\beta$ is accommodated by brittle faulting in the upper crust and by ductile flow in the lower crust and upper mantle.

McKenzie’s uniform stretching model:

Passive pure shear extensional model considers isostatic and thermal consequences of lithosphere stretching.
Passive Rift Model

- As the lithosphere is thinned, it is partly replaced by hotter mantle due to PASSIVE asthenospheric rise to compensate the extension isostatically
- → subsidence/ uplift upon mechanical stretching can be obtained
- The amount of initial (instantaneous) subsidence will depend on the initial amount of stretching (\( \beta \)) due to isostasy

**McKenzie’s uniform stretching model**

**Initial subsidence**

\[
S_i = \frac{y_L \left( \rho_m^* - \rho_c^* \right) y_c \left( 1 - \alpha_v \frac{T_m}{2} y_c y_L \right) - \alpha_v T_m \rho_m^* \left( 1 - 1/\beta \right)}{\rho_m^* \left( 1 - \alpha_v T_m \right) - \rho_c^*}
\]

- \( y_L \) = initial thickness of the lithosphere
- \( y_c \) = initial thickness of the crust
- \( \rho_m^* \) = mantle density (at 0°C)
- \( \rho_c^* \) = crustal density (at 0°C)
- \( \rho_b^* \) = density of the basin infill
- \( \alpha_v \) = thermal expansion coefficient for crust and mantle
- \( T_m \) = temperature of the asthenosphere

**GFZ RWTH Aachen University**
McKenzie’s uniform stretching model

Initial subsidence

\[ S_i = \frac{y_L (\rho_m^* - \rho_s^*) y_L}{\rho_m^* (1 - \alpha_v T_m y_L)} \left( 1 - \frac{\alpha_v T_m y_L}{2} \right) \left( \frac{y_L}{y_c} \right) \frac{(1 - \beta)}{(1 - \beta)} \]

uniform stretching model predicts immediate subsidence if crustal thickness is more than 1/8th of the lithosphere thickness

(For a 100km thick lithosphere this would mean a crust of \(\approx\)12 km)

---

Isostatic and thermal consequences of lithosphere extension

-> cooling/heating changes density:

increase in temperature => increase in volume => decrease in density

-> decrease in temperature => decrease in volume => increase in density

\[ \beta = \frac{l_{stretched}}{l_{initial}} \]

=> Thermal subsidence
Thermal subsidence & Plate Cooling Model

Oceanic plates form by lithospheric cooling.

Depth and Heat Flow of ocean floor

Good approx. till ~70 Ma

\[ d = 2.6 + 0.365 \sqrt{t} \]

Slower for greater ages

\[ d = 5.65 - 2.47e^{-0.3t} \]

\[ q = \frac{510}{\sqrt{t}} \]

\[ q = 48 + 96e^{-t/30} \]
isostatic consequences...Seafloor Depth

\[
\begin{align*}
\text{Column of lithosphere} & \quad \text{Ridge} \\
\int_0^w \rho_w \, dz + \int_0^{w+dz} \rho(z, T) \, dz + \int_0^D \rho_s \, dz & = \int_0^t \rho_w \, dz + \int_s^D \rho(z, T) \, dz \\
\rho(z, T) & = \rho_w \cdot [1 - a(T(z, t) - T_a)] \quad T(z, t) = T_a \cdot \text{erf} \left( \frac{z}{2\sqrt{\kappa \cdot t}} \right)
\end{align*}
\]

Fig. 2.16 Schematic diagram of the cooling oceanic lithosphere at a mid-ocean ridge. The oceanic plate moves away from the ridge at a velocity \( u \). Its age is therefore determined by \( x/u \), where \( x \) is the horizontal distance from the ridge crest.

Fig. 2.17 Calculated isotherms for an oceanic lithosphere that is instantaneously cooled. The values of the isotherms are \( T-T_0 \). The dots are the estimated thicknesses of the oceanic lithosphere in the Pacific, from Levesq et al. (1974).

Fig. 2.18 The principle of isostasy requires the ocean to deepen with age to offset the effects of thermal contraction of the oceanic lithosphere. The water depth below the level of the ridge crest is \( w \), the thickness of the oceanic lithosphere is \( w \), and \( \rho_w, \rho_s \), and \( \rho \) are the mantle, water, and lithospheric densities respectively.

Fig. 2.19 Depths of the ocean floor below the level of the ridge crest as a function of age of the seafloor (Peterson and Schäfer 1971). The solid line shows the theoretical result for an instantaneous cooling model. It is in close agreement with observations from the North Pacific and North Atlantic. The oceanic bathymetry follows a rate-age relationship.

Allen & Allen 2005: Basin Analysis
Gradual cooling (50 - 100 m.y.) & related density increase leads to basin subsidence above and progressive infill with sediment.

**Passive Rift Model**
- McKenzie’s uniform stretching model
- Passive pure shear extensional model

**McKenzie’s uniform stretching model**

Subsidence caused by thermal contraction

\[ S(t) \approx E_0 \frac{\beta}{\pi} \sin \left( \frac{\pi}{\beta} \left(1 - e^{-t/\tau} \right) \right) \]

\[ E_0 = 4y_c \rho^* \alpha_c T_m / \pi^2 (\rho_m^* - \rho_c) \]

Final subsidence

\[ S_v = y_v (1 - 1/\beta) \left[ \frac{\rho_m - \rho_c + \rho_m^* (\alpha_c/2) T_m + \varepsilon}{\rho_m - \rho_w} \right] \]

\[ \varepsilon = \left( \frac{\rho_m^* - \rho_w^*}{\beta} \right) \frac{\alpha_c T_m y_v}{2 y_1} \]
2-phase subsidence

Predicts 2-stage subsidence:

1. $S_t$ as a result of tectonic stretching – on a short time scale, <20 my, and

2. $S_{th}$ as a result of thermal subsidence – long time scale, ca. 50 – 100+ my.

quantifying extension of lithospheric plates

- Where to know $\beta$ from?
extension of lithospheric plates

- Where to know $\beta$ from?

$\rightarrow$ estimate $\beta$ from deep seismic information

Passive Rift Model
McKenzie’s uniform stretching model

A deep reflection seismic line across the Northern Rhine Graben Wund et al., 1991
$\beta$ from preserved present-day crustal thickness

Where to know $\beta$ from?

→ estimate $\beta$ from deep seismic information
→ estimate $\beta$ from tectonic subsidence

subsidence curves

lithospheric mantle re-equilibrated! does not conserve $\beta$...
extension of lithospheric plates

Structurally the uniform stretching model predicts:

**Syn-rift** extensional faults in the upper crust with graben fill overlain by onlapping post-rift sediments of the thermal subsidence phase.

**Passive Rift Model**

McKenzie’s uniform stretching model

Passive pure shear extensional model

---

Magmatic Implications of the Pure Shear Model

For $\beta > 3$ formation of decompressional melts!

- Melts are trapped around the crust-mantle boundary as they are less dense than the lithospheric mantle but denser than the crust.
- **underplating**
- Melts weaken the lithosphere
- **continental breakup**

**pure shear**: temperature of the uprising asthenosphere exceeds the solidus of the mantle and allows melting.
after breakup: magma rich passive margin

What are the observations?

seismic data:
structure and velocity distribution

breakup unconformity below deposits of thermal subsidence

synrift grabens

What are the observations in the sediments...

reflection seismic data: structure

breakup unconform

What are the observations in the crust...?

refraction seismic data: velocity distribution

Comparison of profiles from northeast Greenland and Lofoten-Vesteralen Margin (Voss et al., 2009)

high velocity bodies in the lower crust near COT
seismic exercise
cold (old) magma-rich passive margin

heat flow increase from ocean to continent

breakup unconformity
syn-rift

positive gravity anomaly coinciding spatially with reflective (magmatic?) lower crust

Additional Information
other passive rift models...

the simple shear model

Simple Shear Models

Structurally the simple shear model predicts:

**Syn-rift** extensional faults in the upper crust with **graben fill** detached along listric master faults and offset from overlying **post-rift sediments** of the thermal subsidence phase.
Implications of the Simple Shear Models

Some thermal consequences:

With McKenzie’s pure shear model decompression leads to melting of upper mantle material. => magma-rich passive margins

With Wernicke’s simple shear model it is very difficult to produce sufficient decompression to allow magma formation. => magma-poor passive margins

Implications of the Simple Shear Models

different thermal consequences:

**pure shear:** temperature of the uprising asthenosphere exceeds the solidus of the mantle and allows melting.

**simple shear:** temperature of the uprising asthenosphere never reaches the solidus - so no melting occurs.
if rifting proceeds and leads to breakup how will the evolving passive margin look like?

after breakup: magma poor passive margin

No high velocity/high density bodies in the lower crust, no SDRs
Conjugate Flemish Cap—Galicia Bank margin pair

general crustal structure,

low-angle normal faults, brittle, asymmetric, nonvolcanic, exhumed mantle, low surface heat flow

What are the observations to discriminate between different models?

seismic
gravity
seismology
heat flow
deformation mode
extension of lithospheric plates

- may lead to continental rifting followed by breakup
- depending on the mode and the rate of extension different types of passive margins may develop

- magma rich margins
  - have lots of syn-rift volcanism, high velocity-high-density bodies in the lower crust, SDRs
  - are hot or cold depending on age

- magma-poor margins
  - have very little syn-rift volcanism, uniform lower crust, No SDRs, but large low-angle detachment fault systems
  - are rather cold throughout their history

passive rifting is NOT the only extensional mechanism of basins formation!
**Active Rift Model**

*(+dynamic topography)*

**Active** rise of asthenosphere provokes initial thermal uplift and erosion.

Subsequent cooling leads to subsidence and results in lithosphere thinning =

\[ \frac{d_{\text{initial}}}{d_{\text{stretched}}} \]

---

**Active Rift Model**

predicts

1. **Uplift** as a result of mantle upwelling - short time scale, ca. 10 – 20+ my. (- thermal anomaly)

2. **thermal subsidence** as a result of cooling – long time scale, ca. 50 – 100+ my.
Active Rift Model

what may be left only...

1. An erosional unconformity at the base (Why?)

2. Little preserved faulting and wide-spanned subsidence (sag basin) with onlaps at the margins (bull-head structure)

Isostatically, thinning of the lithospheric mantle results in net surface uplift as heavier lithospheric mantle is replaced by less dense (partially molten) asthenospheric material (active rift model).

Only interaction with surface processes (erosion) and shallow extension due to upward flexure of the plate gives net subsidence due to subsequent cooling of the asthenospheric material.
East African Rift System

EARS=THE active rift model:

high topography, high heat flow, strong volcanism

hot anomaly in asthenospheric mantle below

East African Rift System – active rifting: high topography

www.geology.com/articles/east-africa-rift.shtml
EARS volcanism, + high heat flow

www.geology.com/articles/east-africa-rift.shtml

East African Rift System – active rifting

www.geology.com/articles/east-africa-rift.shtml
Crustal structure of the northern Main Ethiopian Rift from EAGLE

Ethiopia Afar Geoscientific Lithospheric Experiment

(a) Line 1
(b) Line 2

Main Ethiopian Rift from EAGLE

rather uniform crustal velocities
Upper mantle P-wave speed variations beneath Ethiopia and the origin of the Afar hotspot
Margaret H. Benoit et al. 2006

passive seismology to unravel mantle properties
slow anomaly in the mantle below the rift
Take home

- 2 main types of extensional basin forming mechanisms: active vs passive rifting, but several sub-types (pure shear, simple shear, depth-dependent, rheology-dependent),

- deformation (subsidence history) depends on rate of extension and rheological structure of the lithosphere
  1) **magma-poor** (pre-breakup), Andersonian faulting, Wernicke-simple shear model; slow, cold, exhumed & serpentinized mantle at seafloor
     => plate-driven?
     transitional cases...
  2) **magma-rich**: intensive pre-breakup magmatism (LIPs, Plumes or hot spots in der Nähe); SDR, HDB/HVB in der extremely thinned continental crust next to COB; often enigmatic high rift shoulders; hot, fast? breakup
     => mantle-driven?

- McKenzie's uniform streching model describes postrift subsidence amazingly well, but dynamic models considering rheological heterogeneities needed to explain the large variety of different extensional basins and magmatic expressions

more specific models of basin formation...

- consider stratified and heterogenous rheology of the lithosphere

- variations in stretching rate

- predict many different rifting/breakup scenarios
Strength profiles: jelly sandwich versus crème brulée

more sophisticated thermo-mechanical models of basin formation...

Fig. 2.38

Allen & Allen, 2013
meanwhile models go 4D!

plume-lithosphere interaction  
slow and longlasting intracontinental basins

Mechanisms of basins formation

1. extension of lithospheric plates
2. flexure due to vertical or horizontal loads
Basins due to flexure

Flexure = long-wavelength bending of the elastic lithosphere in response to vertical or horizontal loads

wavelength of flexure depends on flexural rigidity of lithosphere and magnitude of load

flexural rigidity of lithosphere is age dependent (old cold strong versus young hot weak)

dependent}=> viscous relaxation = relatively fast process (<10⁵-10⁶ years)
Basins due to flexure

\[ D \frac{d^4 \omega}{dx^4} + \Delta \rho \omega = 0 \]

- \( D \): flexural rigidity
- \( \omega \): deflection
- \( \alpha \): flexural parameter
- \( \rho \): density
- \( x_0 \): half wavelength
- \( x = 0 \): maximum \( \omega \)
- \( x_0 \): \( \omega = 0 \)

Fig. 4.7 (a) Deflection of a continuous elastic plate under a line load. (b) Theoretical deflection of the elastic lithosphere under a line load applied at the centre of an infinitely extensive plate. Parameters are defined in the text. The deflection \( \omega \) is scaled against the maximum deflection \( \omega_0 \). The horizontal distances are scaled against the flexural parameter \( \alpha \). Bending moments and horizontal in-plane forces are zero. From Turcotte & Schubert (2002); © Cambridge University Press, 2002.

Fig. 4.1 Flexure at sites of plate convergence. Schematic illustration of (a) pairs of pre-foreland and retro-foreland basins at sites of continent–continent collision, as in the Alps; (b) retro-foreland basins and ocean trenches at sites of ocean–continent convergence, as in the Andes (especially the Chicamal toco) and Severn belt of the USA, where subduction is at an active ocean; and (c) pre-foreland basins and backarc extension related to subduction zone rollback, as in the Apennine–Tethysian chain, and simple subduction of the oceanic plate, as in the Marianas. Based partly on Lydeard & Karumori (1979) and Stern (2002).
Fig. 4.19 Different styles of periodic buckling by a compressional force \( f \); \( h_c \) and \( h_m \) are the thicknesses of the competent crust and mantle respectively, which fold with wavelengths \( \lambda_c \) and \( \lambda_m \). Where the lower crust is very weak, the upper crust and mantle lithosphere fold with different wavelengths (\( \lambda_c < \lambda_m \)), producing decoupled or biharmonic folding (b). Where there is a single coherent layer of coupled crust and mantle, as in both young (<130 Ma) and very old (>1000 Ma) lithospheres, monoharmonic folding develops (a and c). After Burov et al. (1995) and Cloetingh et al. (1999).

**Folded lithosphere of Iberia**

Topography displaying cylindrical patterns of alternating parallel trending highs and lows and corresponding gravity anomalies

analogue modelling experiments, displaying pop-up structures underlying topographic highs, accompanied by folding of the Moho (Fernandez-Lozano et al., 2010)
the influence of the deep mantle..

dynamic topography (+/-)

positive dynamic topography
Uplift & Erosion
rising hot/low-density material in the asthenosphere
cooling & subsidence
sub-type: active rift model

negative dynamic topography
subsidence
sinking cold/high-density material in the asthenosphere (why?)
Subsidence history similar to lithosphere folding

Heat Flow
shallow - hot
deep - cold
“Transtensional”, pull-apart basins, e.g., Salton Sea, Dead Sea etc.
orientation of principal stress axes in strike-slip regime?

**Sigma 1 horizontal**

**Sigma 2 vertical**

**Sigma 3 horizontal**

DIFFERENT from extensional basins

- **Sigma 1 vertical**
- **Sigma 2 horizontal**
- **Sigma 3 horizontal**

DIFFERENT from compressive basins

- **Sigma 1 horizontal**
- **Sigma 2 horizontal**
- **Sigma 3 vertical**

**strike-slip basins**
Transform? Transcurrent?

Transform --- plate (lithosphere-) scale e.g. San Andreas F
Transcurrent --- crustal scale e.g. Garlock Fault CA

(USGS)

strike-slip basin types

- narrow (km-10-nm, elongated (up to 1000 km)
- asymmetry, abrupt facies changes
- negative/positive flower structures
- Sediments above steep basement fault
- fast, short-lived subsidence
- high sedimentation rates (1m/1000a)
- lateral migration of depo centres

Nissen and Sylvester, 1996
how to find out that its not a rift?
Strike-slip PDZ in Cross-section

- Flower-structures
- Variable offsets along the same fault because of movement in and out of the plane of observation

Figure 1. Generic three-stage model for shear margin formation (after Lorenzo, 1997): (1) rift: continent-continent shearing; (2) drift: continent-ocean transform boundary (active margin); and (3) passive margin: continent-ocean fracture zone boundary.
Sheared Margins Gulf of Guinea

Early Cretaceous, 125 Ma: Rifting initiated.

Late Albian, 100 Ma: Final contact between the continental crusts of Brazil and Africa reached.

Aptian, 110 Ma: Smaller divergent basins created. Rapidly deposited clastics are deformed along shear zones.

Upper Cretaceous (Turonian-Oxfordian times), 65 Ma: Correlation between Central Atlantic and South Atlantic oceanic basins established.

---

crustal structure @ Sheared Margins

N-S reflection seismic line across the Côte d’Ivoire-Ghana transform margin, Basile et al., 1993
To sum up...

Typical subsidence histories

Plate tectonics and basin subsidence history
Xiangyang Xie and Paul L. Heller
GSA Bulletin; January 2009; v. 121; no. 1-2; p. 55-64; DOI: 10.1130/B26398.1
typical tectonic subsidence rates

Allen & Allen, 2013, Fig. 9.17

Different thermal histories

depending on
mechanism
rate
rheology

very characteristic
thermal and subsidence histories
defformation pattern (brittle vs ductile)
factors influencing the thermal field

- basal heat flow (thinning mechanism, extensional/ flexural?)
- age (time since basin initiation, depth of thermal LAB)
- sedimentation rate
- lithology of basin fill and basement (densities, thermal conductivities & radiogenic heat production)
- compaction history (highly compacted -> dense, highly conductive)
- crystalline basement structure