

# Cambridge Part III Maths

Lent 2020

## Fluid Dynamics of the Solid Earth

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Notes created based on Josh Kirklin's L<sup>A</sup>T<sub>E</sub>X packages & classes. Please do not distribute these notes other than to fellow Part III students. Please send errors and suggestions to [cwp29@cam.ac.uk](mailto:cwp29@cam.ac.uk).

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Lecture 1  
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## 1 Introduction

The course will use the wealth of observations of the solid Earth to motivate mathematical models of the physical processes governing its evolution. The dynamic evolution is governed by a rich variety of physical processes occurring on a wide range of length and time scales.

- The Earth’s core is formed by the solidification of a mixture of molten iron and various light elements, a process which drives predominantly compositional convection in the liquid outer core, thus producing the geodynamo responsible for the Earth’s magnetic field.
- On million year timescales, the solid mantle convects, and as it upwells to the surface it partially melts leading to volcanism.
- At the surface, convection drives the motion of brittle plates which are responsible for the Earth’s topography as can be felt and imaged through the seismic record (figure 1)
- In the Earth’s surface, fluids flow through porous rocks, for example groundwater aquifers which feed streams and rivers which erode the solid surface.
- On the Earth’s surface, similar physical processes of viscous and elastic deformation coupled to phase changes govern the evolution of the Earth’s cryosphere, from the solidification of sea ice to the flow of glacial ice over land and ice shelves over the ocean.

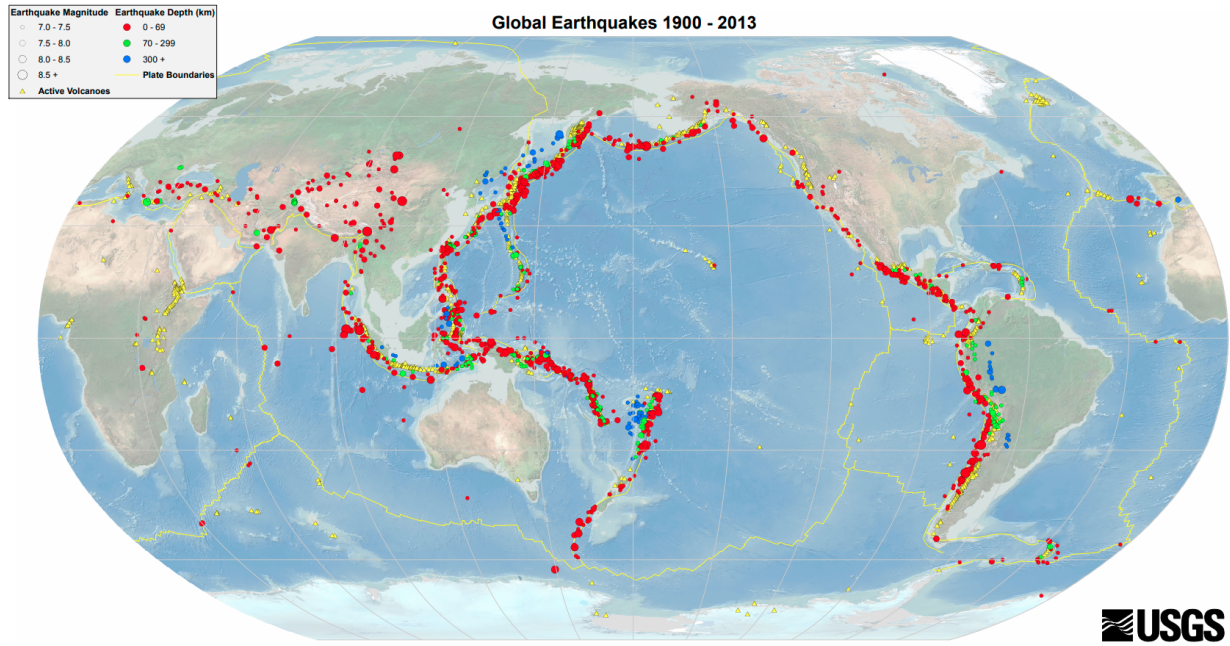


Figure 1: Map of global earthquakes, visibly localised to tectonic plate boundaries.

Predominantly, the mathematics is of slow viscous flows. Topics include the onset and scaling of convection, the coupling of fluid motions with changes of phase at a boundary, the thermodynamic and mechanical evolution of multicomponent or multiphase systems, the coupling of fluid flow and elastic flexure or deformation, and the flow of fluids through porous materials.

## 2 Plate cooling

Here we consider a half-space cooling model of the oceanic lithosphere (oceanic plates). The bottom surface of Earth's oceans, particularly clear in the Atlantic ocean, has a large scale structure in which the middle of the ocean (the *mid-ocean ridge*) is shallower than regions closer to the continents. The mid-ocean ridge forms as a result of separating tectonic plates. We know that the plates move apart here due to magnetic anomalies forming 'stripes' of alternating polarity. The quasi-periodic flipping of the Earth's magnetic polarity allows dating of the stripes. The plates are driven apart by convection of the Earth's mantle.

### 2.1 Thermal problem

We wish to form a model describing the depth of the ocean floor near mid-ocean ridges. First we estimate the temperature field. Consider an idealised model with a flat surface (for now), observed plate spreading rate  $U$ , surface temperature  $T_0$ , and deep mantle temperature  $T_1$ . The temperature field is described by the advection-diffusion equation

$$\rho c_p \left( \frac{\partial T}{\partial t} + \mathbf{u} \cdot \nabla T \right) = \nabla \cdot (k \nabla T)$$

where  $c_p$  is specific heat capacity,  $k$  is thermal conductivity,  $\rho$  is density, all assumed constant. For simplicity, we combine these constants into the thermal diffusivity  $\kappa = k/\rho c_p$ . Then

$$\frac{\partial T}{\partial t} + \mathbf{u} \cdot \nabla T = \kappa \nabla^2 T$$

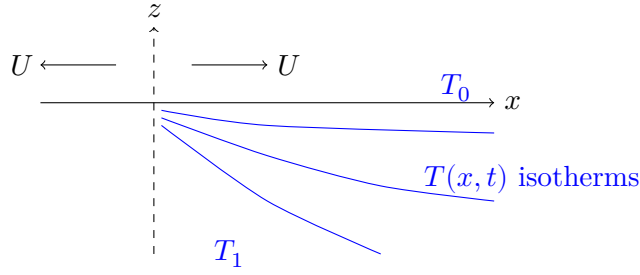


Figure 2: Schematic diagram of mid-ocean ridge spreading and mantle temperature isotherms.

We wish to find the steady state profile with  $\partial_t = 0$ ,  $\mathbf{u} = U\hat{\mathbf{x}}$  where  $U$  is constant. Note that far from the ridge axis, the thickness of the plate is much smaller than the extent of the plate. Hence in terms of scalings,  $z \ll x$  and we may neglect the  $\partial_x^2$  component of  $\nabla^2$ . We have

$$U \frac{\partial T}{\partial x} = \kappa \left( \frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial z^2} \right) \approx \kappa \frac{\partial^2 T}{\partial z^2} \quad (1)$$

The scaling given by this equation is  $U\Delta T/x \sim \kappa\Delta T/z^2$  where  $\Delta T = T_1 - T_0$  is the natural temperature scale. There is no natural lengthscale, so we use that given by the advection-diffusion equation:

$$z \sim \sqrt{\frac{\kappa x}{U}}$$

We can proceed by finding a self-similar solution with similarity variable

$$\eta = \frac{z}{2\sqrt{\frac{\kappa x}{U}}}$$

and seek solutions of the form

$$\theta = \frac{T - T_0}{T_1 - T_0} = \theta(\eta)$$

Using the variables  $\eta, \theta$ , (1) becomes

$$\begin{aligned} -U\Delta T \frac{\eta}{2x} \theta_\eta &= \frac{\kappa\Delta T}{4\frac{\kappa x}{U}} \theta_{\eta\eta} \\ \Rightarrow \theta_{\eta\eta} + 2\eta\theta_\eta &= 0 \end{aligned}$$

We can integrate directly to get  $\theta_\eta = ae^{-\eta^2}$ , which gives

$$\theta = b + a \int_0^\eta e^{-y^2} dy$$

The boundary conditions are  $\theta(0) = 1$  and  $\theta(\infty) = 0$  based on the definitions of  $T_0$  and  $T_1$ . The thermal structure away from the ridge is then

$$T = T_0 + (T_1 - T_0) \operatorname{erf} \left( \frac{z}{2\sqrt{\frac{\kappa x}{U}}} \right) \quad (2)$$

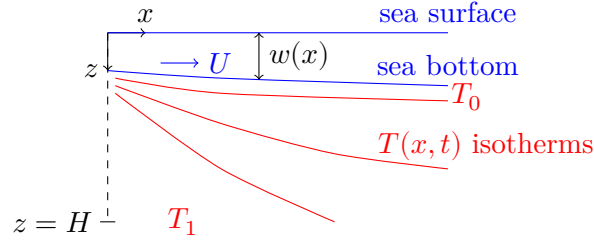


Figure 3: Schematic diagram of ocean depth and crust temperature surfaces.

where the error function  $\text{erf}$  and its complement  $\text{erfc}$  are defined by

$$\text{erf}(x) = \frac{2}{\sqrt{\pi}} \int_0^x e^{-y^2} dy$$

$$\text{erfc}(x) = 1 - \text{erf}(x)$$

## 2.2 Ocean depth away from mid-ocean ridge

We now consider the depth of the ocean following from the temperature field derived above. We choose axes with  $z$  increasing downwards, placing the sea surface at  $z = 0$  and the ‘bottom’ of the mantle at  $z = H$ . The coordinate  $x$  increases away from the mid-ocean ridge, with ocean depth  $w(x)$  at a given  $x$ .

First, consider *isostasy*: Archimedean buoyancy applied to Earth’s crust. This indicates the depth at which an object/fluid parcel of some density lies in a fluid of different density.



Denoting the density of the crust and mantle as  $\rho_c, \rho_m$  respectively, the crust of thickness  $h$  sits at a depth  $b$  in the mantle. Hydrostatic balance gives  $\rho_c g h = \rho_m g b$ . Equivalently, we can consider a force balance between the weight of the crust and the buoyancy force:

$$\rho_c (h - b) g = (\rho_m - \rho_c) g b$$

Within the oceanic lithosphere we have a density field

$$\rho = \rho_m (1 - \alpha(T - T_1))$$

where  $T = T(x, z)$  is the thermal model derived above, given by (2). Isostatic balance gives the following, which balances water weight and mantle weight, with water and mantle buoyancy. The ocean density is denoted by  $\rho_w$ .

$$\begin{aligned}
 \rho_w w_0 + \rho_m (H - w_0) &= \rho_w w(x) + \int_w^H \rho(T) dz \\
 &= \rho_w w + \rho_m (H - w) - \rho_m \alpha \int_w^H (T - T_1) dz \\
 \Rightarrow (\rho_m - \rho_w)(w - w_0) &= -\rho_m \alpha \int_w^H (T - T_1) dz \\
 &\approx -\rho_m \alpha (T_1 - T_0) \int_0^\infty \text{erfc}(\eta) \cdot 2\sqrt{\frac{\kappa x}{U}} d\eta
 \end{aligned}$$

where the last approximate equality follows from taking  $w \rightarrow 0$  and  $H \rightarrow \infty$ , approximating the fact the mantle is much deeper than the ocean. The ocean depth is therefore

$$w - w_0 = \frac{\rho_m}{\rho_m - \rho_w} \alpha (T_1 - T_0) \frac{2}{\sqrt{\pi}} \left( \frac{\kappa x}{U} \right)^{1/2}$$

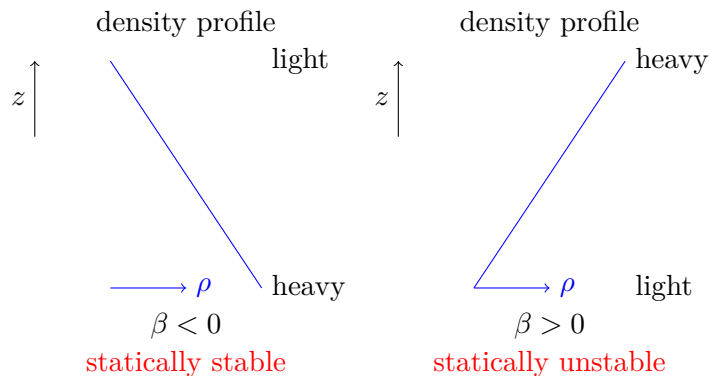
This model fits the data well near to the mid-ocean ridge, with crust age up to 75 million years. However, away from the ridge, the model breaks down as  $w$  tends to a constant as  $x \rightarrow \infty$ . The breakdown of the model is due to convection: the infinite depth mantle approximation breaks down and convection dynamics become important.

### 3 Natural convection

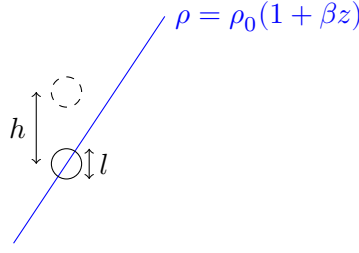
Natural convection arises in flows driven by density differences in a gravitational field, e.g. due to temperature or composition.

#### 3.1 Static stability

Consider the case of no fluid motion  $\mathbf{u} = 0$  and initial stratification  $\rho = \rho_0(1 + \beta z)$ . If  $\beta < 0$ , the dynamics are statically stable. If  $\beta > 0$ , the dynamics are statically unstable but dynamically could be stable.



**Scaling analysis.** Consider a fluid parcel of characteristic size  $l$  in unstable density profile  $\rho = \rho_0(1 + \beta z)$ ,  $\beta > 0$ . Suppose the parcel moves up a distance  $h$  in time  $\tau$ .



The rise of the fluid parcel releases potential energy  $E$  which scales as

$$E \sim (\Delta\rho l^3)gh \sim (\rho_0\beta hl^3)gh \sim \rho_0\beta gh^2l^3$$

The timescale for the rising motion is limited by diffusion of buoyancy (i.e. diffusion of temperature in this case) so  $\tau \sim l^2/\kappa$  where  $\kappa$  is thermal diffusivity. Viscous dissipation over timescale  $\tau$  scales as the shear stress times the distance travelled:

$$\mathcal{D} \sim \frac{\mu U}{l} l^2 h \sim \mu \frac{h/\tau}{l} l^2 h \sim \frac{\mu \kappa h^2}{l}$$

Instability arises if  $E \gtrsim \mathcal{D}$ , which we can write as

$$\rho_0\beta gh^2l^3 \gtrsim \frac{\mu \kappa h^2}{l}$$

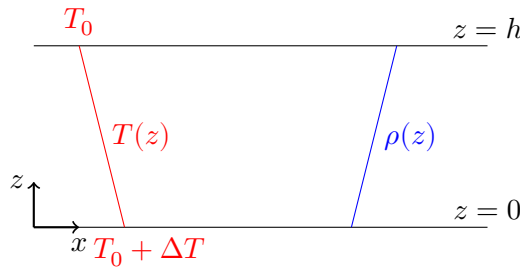
Define the *Rayleigh number*

$$\text{Ra} = \frac{\rho_0\beta gl^4}{\kappa\mu}$$

which quantifies the ratio of timescales of thermal transport via diffusion versus via convection. If Ra is large, the flow is more turbulent. There is instability if  $\text{Ra} \gtrsim \mathcal{O}(1)$ .

### 3.2 Onset of convection

Consider a fluid of depth  $h$  between  $z = 0$  and  $z = h$  with density profile  $\rho(z)$  and temperature difference  $\Delta T$  across the depth, where the  $z = 0$  surface is warmer.



The governing equations are conservation of mass (3), conservation of momentum (4), and conservation of (thermal) energy (5).

$$\nabla \cdot \mathbf{u} = 0 \tag{3}$$

$$\rho \left( \frac{\partial \mathbf{u}}{\partial t} + \mathbf{u} \cdot \nabla \mathbf{u} \right) = -\nabla p + \mu \nabla^2 \mathbf{u} - \rho g \hat{\mathbf{z}} \tag{4}$$

$$\rho c_p \left( \frac{\partial T}{\partial t} + \mathbf{u} \cdot \nabla T \right) = \nabla \cdot (k \nabla T) \tag{5}$$

For steady solutions with  $\mathbf{u} = 0$  and  $\partial_t = 0$  we find

$$\begin{aligned} T &= T_0 + \Delta T \left(1 - \frac{z}{h}\right) \\ \frac{\partial p}{\partial x} &= \frac{\partial p}{\partial y} = 0 \\ \frac{\partial p}{\partial z} &= -\rho_0 g (1 - \alpha(T - T_0)) \end{aligned}$$

We assess the stability by examining small perturbations to a steady base state:

$$\begin{aligned} \mathbf{u} &= 0 + \mathbf{u}'(\mathbf{x}, t) \\ T &= T_0 + \Delta T \left(1 - \frac{z}{h}\right) + T'(\mathbf{x}, t) \\ p &= p_0(z) + p'(\mathbf{x}, t) \end{aligned}$$

The linearised perturbation equations are thus

$$\begin{aligned} \nabla \cdot \mathbf{u}' &= 0 \\ \rho_0 \frac{\partial \mathbf{u}'}{\partial t} &= -\nabla p' + \mu \nabla^2 \mathbf{u}' + \rho_0 g \alpha T' \hat{\mathbf{z}} \\ \frac{\partial T'}{\partial t} - \frac{\Delta T}{h} w' &= \kappa \nabla^2 T' \end{aligned}$$

**Non-dimensionalising.** We wish to non-dimensionalise these equations. There are two intrinsic scales, temperature  $\sim \Delta T$  and lengths  $\sim h$ . We form velocity and time characteristic scales via *diffusive scaling*. From the thermal equation:

$$\frac{\Delta T}{t} \sim \frac{\Delta T U}{h} \sim \frac{\kappa \Delta T}{h^2}$$

From the second relation we have  $U \sim \kappa/h$ , then from the first we have  $t \sim h^2/\kappa$ . The non-dimensionalised equations (dropping ' henceforth) are

$$\begin{aligned} \nabla \cdot \mathbf{u} &= 0 \\ \frac{1}{\text{Pr}} \frac{\partial \mathbf{u}}{\partial t} &= -\nabla p + \nabla^2 \mathbf{u} + \text{Ra} T \hat{\mathbf{z}} \\ \frac{\partial T}{\partial t} - w &= \nabla^2 T \end{aligned} \tag{6}$$

where the *Prandtl number* is defined as  $\text{Pr} = \frac{\mu/\rho_0}{\kappa}$  which quantifies the importance of viscous diffusion versus thermal diffusion.

**Boundary conditions.** We require boundary conditions on the temperature  $T$  and velocity  $\mathbf{u}$ . There are several possible conditions, which may be mixed:

- Rigid boundary  $[\mathbf{u} \cdot \hat{\mathbf{n}}]_{\text{boundary}} = 0$  where  $\hat{\mathbf{n}}$  is outwards normal of boundary surface
- No-slip  $[\mathbf{u} \times \hat{\mathbf{n}}]_{\text{boundary}} = 0$
- Stress-free (free slip)  $\mu \frac{\partial u}{\partial n} = \mu \frac{\partial v}{\partial n} = 0$  i.e.  $n = z$  in our picture
- Fixed temperature  $T_{\text{boundary}} = \text{const.}$



- Fixed heat flux  $[\hat{\mathbf{n}} \cdot \nabla T]_{\text{boundary}} = \text{const.}$

We eliminate pressure from the non-dimensionalised equations by taking  $\hat{\mathbf{z}} \cdot \nabla \times (\nabla \times (6))$ . Recall the vector identity

$$\nabla \times (\nabla \times \mathbf{A}) = \nabla(\nabla \cdot \mathbf{A}) - \nabla^2 \mathbf{A}$$

The governing equations are reduced to

$$(\text{Pr}^{-1} \partial_t - \nabla^2) \nabla^2 w = \text{Ra} \nabla_H^2 T \quad (7)$$

$$(\partial_t - \nabla^2) T = w \quad (8)$$

where  $\nabla_H = (\partial_x, \partial_y, 0)$  is the horizontal gradient.

**Normal modes.** We examine growth/decay of perturbations of the form

$$(w, T) = (\hat{w}(z), \hat{T}(z)) e^{i(kx + ly) + \sigma t}$$

For convenience, and given the  $x, y$  symmetry, define  $a^2 = k^2 + l^2$  as an effective horizontal wavenumber. We have from the governing equations (7), (8):

$$[\sigma \text{Pr}^{-1} - (D^2 - a^2)] (D^2 - a^2) = -\text{Ra} a^2 \hat{T} \quad (9)$$

$$[\sigma - (D^2 - a^2)] \hat{T} = \hat{w} \quad (10)$$

where  $D = d/dz$ . The boundary in our case will be rigid, stress free, and fixed temperature. Note that this is not necessarily representative of the actual boundary conditions, but is useful analytically. Thus at  $z = 0, 1$  we have  $w = 0$  (rigid),  $D^2 w = 0$  (stress free) and  $T = 0$  (fixed temperature). Given these conditions, the heat equation implies  $D^2 T = 0$  on  $z = 0, 1$  and the momentum equation implies  $D^4 w = 0$  on  $z = 0, 1$ .

**Marginal stability.** The marginal stability boundary is  $\sigma = 0$  between growth ( $\sigma > 0$ ) and decay ( $\sigma < 0$ ). Combining (9) and (10) with  $\sigma = 0$  we have

$$(D^2 - a^2)^3 (\hat{T}, \hat{w}) = -a^2 \text{Ra} (\hat{T}, \hat{w})$$

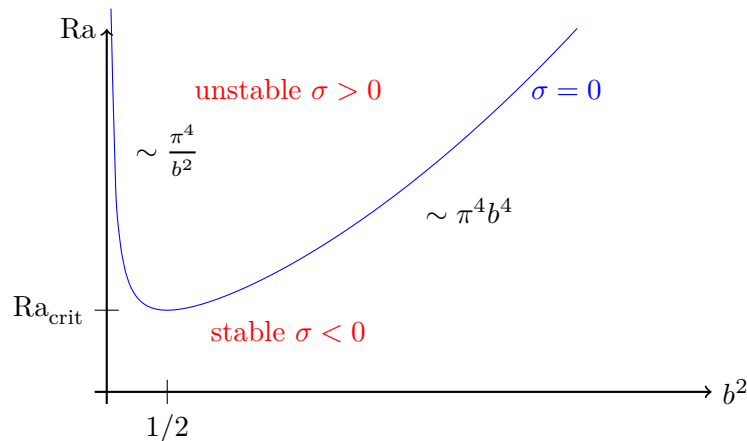
The solutions have structure

$$(\hat{T}, \hat{w}) = (\hat{T}_0, \hat{w}_0) \sin n\pi z$$

Hence we have the eigenvalue condition  $(n^2 \pi^2 + a^2)^3 = a^2 \text{Ra}$ . Let  $a^2 = \pi^2 b^2$ . Then

$$\text{Ra} = \frac{(n^2 + b^2)^3 \pi^4}{b^2}$$

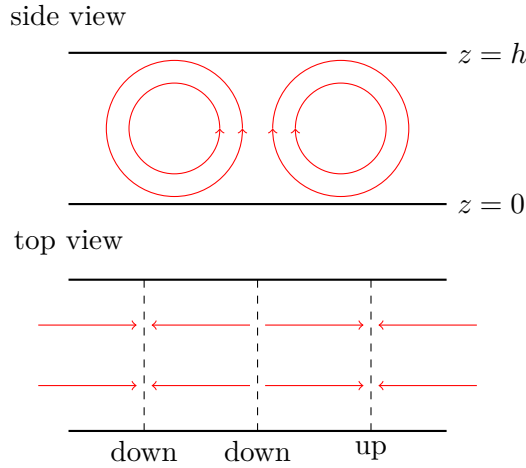
The minimum Rayleigh number is at  $n = 1$  where  $\text{Ra} = (1 + b^2)^3 \pi^4 / b^2$ :



Setting  $\partial \text{Ra} / \partial b^2 = 0$  to find the critical Rayleigh number  $\text{Ra}_{\text{crit}}$ , we find

$$\text{Ra}_{\text{crit}} = \frac{27\pi^4}{4} \approx 657.5 \quad \text{at } a = \frac{\pi}{\sqrt{2}}$$

Hence the dimensional wavelength at critical Rayleigh number is  $2\pi h/a = 2\sqrt{2}h \approx 2.8h$ . The convection process appears like so:



**Example. Mantle convection.** In the mantle, internal heating is due to radiogenic decay. We model this by adding a source term to the conservation of energy equation:

$$\rho_0 c_p \left( \frac{\partial T}{\partial t} + \mathbf{u} \cdot \nabla T \right) = \nabla \cdot (k \nabla T) + \rho_0 Q$$

where  $Q = 10^{11} \text{Wkg}^{-1}$ . The thermal boundary conditions are  $T(0) = T_s$ , i.e. fixed surface temperature, and  $\partial T / \partial z = 0$  at  $z = h$ , i.e. zero basal heat flux.

A scaling analysis gives

$$x \sim h, \frac{k \Delta T}{h^2} \sim \rho_0 Q \Rightarrow \Delta T \sim \frac{\rho_0 Q h^2}{k}$$

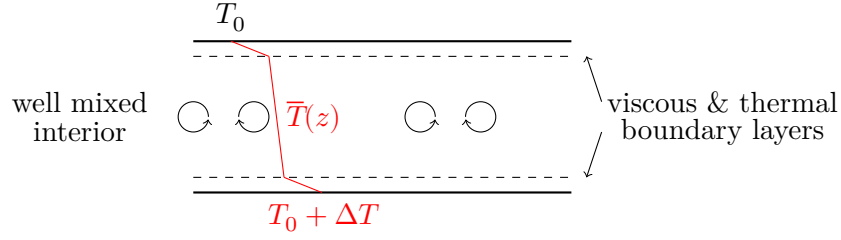
The Rayleigh number in this case is

$$\text{Ra}_Q = \frac{\rho_0 g \alpha (\rho_0 Q h^2 / k) h^3}{\kappa \mu} = \frac{\rho_0^2 g \alpha Q h^5}{k \kappa \mu}$$

Typical values are  $\text{Ra}_Q = 3 \times 10^9$  and  $\text{Pr} = 10^{22}$ . Note that  $\text{Ra}_Q$  is much larger than  $\text{Ra}_{\text{crit}}$ .

### 3.3 High Rayleigh number convection

At high Rayleigh number, the key idea is: small plumes are generated at boundaries, which only ‘see’ the statistically well-mixed interior temperature. Hence heat flux is independent of depth of domain.



The heat flux is characterised by the *Nusselt number* defined by

$$\text{Nu} = \frac{F_h}{k\Delta T/h} = f(\text{Ra}, \text{Pr})$$

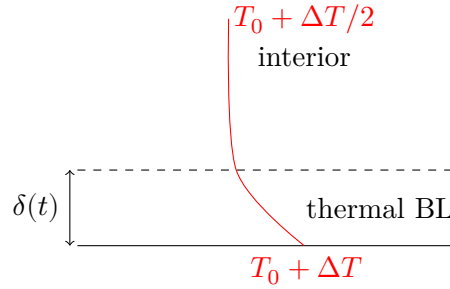
where  $F_h$  is the heat flux, and Nu is a function of Ra, Pr only. If  $F_h$  is not a function of  $h$ , then we must have

$$\begin{aligned} F_h &= \frac{k\Delta T}{h} f(\text{Ra}, \text{Pr}) \sim \frac{k\Delta T}{h} \text{Ra}^{1/3} \\ \Rightarrow F_h &= \lambda(\text{Pr}) k \left( \frac{\rho_0 g \alpha}{K \mu} \right)^{1/3} \Delta T^{4/3} \end{aligned}$$

for some function  $\lambda$ .

**Boundary layer analysis.** The picture was formalised by Lou Howard (1965). We assume the thermal boundary layer grows diffusively (i.e. slowly) according to

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2}$$



Diffusive scaling gives  $z \sim \sqrt{\kappa t}$ , i.e. a self-similar solution with similarity variable  $\eta = z/2\sqrt{\kappa t}$ . The thermal problem can be solved to get

$$T = T_0 + \frac{\Delta T}{2} \left[ 1 + \text{erfc}\left(\frac{z}{2\sqrt{\kappa t}}\right) \right]$$

The boundary layer is unstable when the local Rayleigh number exceeds the critical value.

$$\begin{aligned} \text{Ra}_{BL} &= \frac{\rho_0 g \alpha (\Delta T/2) \delta(t)^3}{\kappa \mu} \gtrapprox \text{Ra}_{\text{crit}} \\ \Rightarrow \frac{\delta}{h} &\gtrapprox \left( \frac{2\text{Ra}_c}{\text{Ra}} \right)^{1/3} \end{aligned}$$

Assume that advection of plumes from the boundary layer is fast compared with diffusive growth of the boundary layer. The heat flux is

$$F_h = \frac{k\Delta T}{2t_c} \int_0^{t_c} \frac{dt}{\sqrt{\pi\kappa t}} = \frac{k\Delta T}{\delta_c}$$

Hence the Nusselt number can be written

$$\text{Nu} = \frac{F_h}{k\Delta T/h} = \frac{h}{\delta_c} = \left( \frac{\text{Ra}}{2\text{Ra}_c} \right)^{1/3}$$

For  $\text{Ra}_{\text{crit}} \approx 1100$ , the Nusselt number is thus  $\text{Nu} \approx 0.077\text{Ra}^{1/3}$  which matches well with experiment.

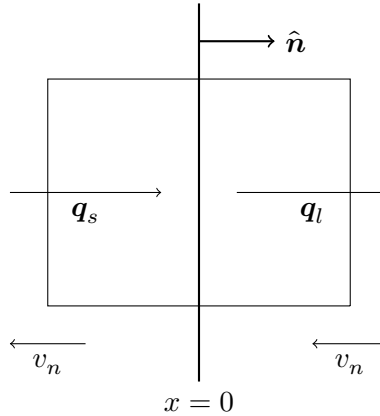
## 4 Solidification/melting

. Applications of solidification/melting theory include

- Formation of Earth's crust (early 1800s)
- Sea ice, due to Stefan (late 1800s)
- Earth's core (mid 1950s)
- Igneous rocks (mid 1980s)

### 4.1 The Stefan condition

Consider a solid–liquid interface  $x = a(t)$  with boundary velocity  $v_n$ . We will work in the frame of reference moving with the boundary.



We assume there is a heat flux  $\mathbf{q}_s$  into the control volume, and a heat flux  $\mathbf{q}_l$  leaving the control volume. Assume equal densities for the liquid and solid, i.e.  $\rho_l = \rho_s = \rho$ . Conservation of energy in the control volume gives:

$$\rho H_l v_n - \hat{\mathbf{n}} \cdot \mathbf{q}_l - \rho H_s v_n + \hat{\mathbf{n}} \cdot \mathbf{q}_s = 0$$

where  $\mathbf{q} = -k\nabla T$  and  $H$  is the *specific enthalpy*, which is the heat energy at constant pressure per unit mass. Rearranging, we have

$$\rho(H_l - H_s)v_n = \hat{\mathbf{n}} \cdot (\mathbf{q}_l - \mathbf{q}_s)$$

Define  $L = H_l - H_s = T_m(s_s - s_l)$  where  $T_m$  is the melting temperature and  $s$  is the entropy of solid/liquid.  $L$  is the latent heat of fusion/solidification. At the boundary  $x = a(t)$  we have

$$\rho L v_n = k \hat{\mathbf{n}} \cdot (\nabla T|_s - \nabla T|_l)$$

where all quantities are evaluated at the boundary. This is the *Stefan condition*.

Note that if  $\rho_s \neq \rho_l$  there is a velocity  $u_n$  induced at the interface:

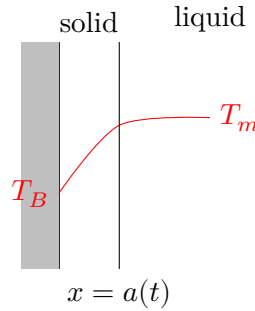
$$u_n = \frac{\rho_l - \rho_s}{\rho_l} v_n$$

in which case the Stefan condition is modified to

$$\rho_s L v_n = k_s \hat{\mathbf{n}} \cdot \nabla T|_s - k_l \hat{\mathbf{n}} \cdot \nabla T|_l$$

## 4.2 1D solification from a cooled boundary

For convenience we again assume equal densities and thermal properties.



In the solid, the temperature  $T$  obeys

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial x^2}$$

with boundary conditions  $T = T_B$  at  $x = 0$  and  $T = T_m$  at  $x = a(t)$ . We also have the Stefan condition

$$\rho L \dot{a} = k \left. \frac{\partial T}{\partial x} \right|_{a^-}$$

at  $x = a(t)$ . Scaling from the thermal equation gives  $x \sim \sqrt{\kappa t}$  as usual. From Stefan's condition we have

$$\rho L \frac{x}{t} \sim k \frac{\Delta T}{x} = \rho c_p \kappa \frac{\Delta T}{x}$$

which yields  $x \sim \sqrt{\kappa t \frac{c_p \Delta T}{L}}$ . The factor in the square root is dimensionless, hence we have  $x \sim \sqrt{\kappa t}$  in agreement with the thermal equation. We use similarity variable  $\eta = x/2\sqrt{\kappa t}$  and define  $\theta(\eta)$  via

$$T - T_B = (T_m - T_B) \theta(\eta)$$

Now  $a(t) = 2\lambda\sqrt{\kappa t}$ , thus the diffusion equation becomes

$$-2\eta\theta' = \theta''$$

with  $0 \leq \theta \leq \lambda$ . The boundary conditions are now  $\theta(0) = 0, \theta(1) = \lambda$  and

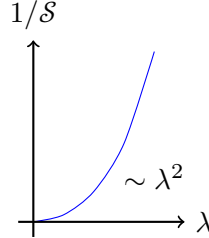
$$2\mathcal{S}\lambda = \theta'(\lambda)$$

where  $\mathcal{S} = L/c_p \Delta T$  is the Stefan number. The solution is

$$\theta = \frac{\operatorname{erf} \eta}{\operatorname{erf} \lambda}$$

where  $\lambda$  satisfies

$$\sqrt{\pi} \lambda e^{\lambda^2} \operatorname{erf} \lambda = \frac{1}{\mathcal{S}}$$



Large  $\mathcal{S}$  implies small  $\lambda$ , i.e. small growth, and vice versa. Note for water,  $L/c_p \approx 80^\circ\text{C}$ .

#### 4.2.1 Quasi-steady approximation

When  $\mathcal{S} \gg 1$  the Stefan condition implies growth is slow. Rescale time  $t \rightarrow \mathcal{S}t$  so the time derivative becomes  $\mathcal{O}(\mathcal{S}^{-1})$ . Then

$$\frac{\partial^2 T}{\partial x^2} \approx 0 \implies \frac{\partial T}{\partial x} = \frac{T_m - T_B}{a}$$

i.e. there is a linear conduction profile in the solid. The Stefan condition is now

$$\rho L \dot{a} = k \left. \frac{\partial T}{\partial x} \right|_a \approx \frac{k(T_m - T_B)}{a}$$

where  $a = \sqrt{2\kappa t/\mathcal{S}}$ . This definition of  $a$  follows from:  $\mathcal{S} \gg 1 \implies \lambda \ll 1$ . Then

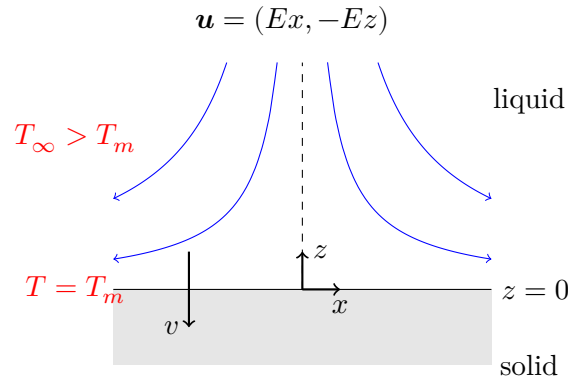
$$\sqrt{\pi} \lambda e^{\lambda^2} \operatorname{erf} \lambda = \frac{1}{\mathcal{S}} \approx \sqrt{\pi} \lambda \cdot 1 \cdot \frac{2}{\sqrt{\pi}} \lambda$$

Hence  $\lambda \sim (1/2\mathcal{S})^{1/2}$ , from which  $a(t) = 2\lambda\sqrt{\kappa t}$  gives  $a$  as above.

Note: as  $t \rightarrow 0$ ,  $\dot{a} \sim t^{-1/2} \rightarrow \infty$ . This is physically implausible; in fact growth is regularised by crystal kinetics which is not included in this model.

#### 4.3 Phase change and fluid flow

Consider 2D stagnation point flow driving a steady melting of a solid, with fluid velocity



Steady-state in a translating frame of reference (i.e. moving with boundary at speed  $v$ ) implies  $T = T(z)$  is a function of  $z$  only. In steady-state are left with a balance of advection and diffusion:

$$(v - Ez) \frac{\partial T}{\partial z} = \kappa \frac{\partial^2 T}{\partial z^2}$$

where  $v$  is the melt rate. Neglecting advection by the frame (justify later) we have solution

$$T = T_m + (T_\infty - T_m) \operatorname{erf} \left( \sqrt{\frac{E}{2\kappa}} z \right)$$

The Stefan condition gives

$$-\rho L v = -k \left. \frac{\partial T}{\partial z} \right|_{z=0} = -k \frac{2}{\sqrt{\pi}} (T_\infty - T_m) \sqrt{\frac{E}{2\kappa}}$$

Hence the melt rate can be written as

$$v = \frac{1}{\mathcal{S}} \sqrt{\frac{2E\kappa}{\pi}}$$

Note the following:

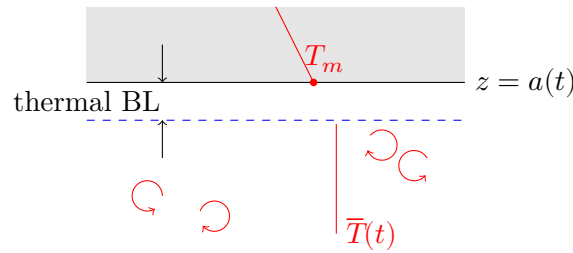
- The melt rate  $v$  increases as flow rate  $E$  increases ( $v \sim \sqrt{E}$ )
- The thermal boundary layer has scale  $\delta_T \sim \sqrt{\kappa/E}$ . Thus  $\delta_T$  decreases as  $E$  increases, i.e. the flow compresses the thermal boundary layer
- The ratio of frame advection to flow compression is

$$\frac{\text{advection}}{\text{compression}} \sim \frac{v}{Ez} \sim \frac{v}{E\delta_T} \sim \frac{v}{\sqrt{E\kappa}} \sim \frac{1}{\mathcal{S}}$$

So neglect of frame advection is valid if  $\mathcal{S} \gg 1$ .

#### 4.4 Solidification with vigorous convection

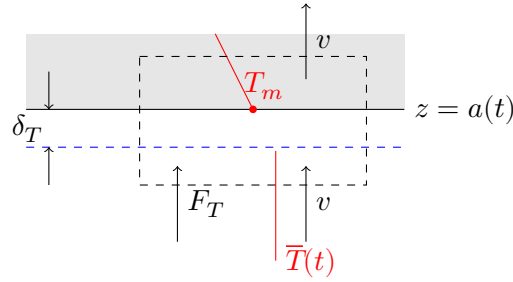
Consider solidification of a liquid with vigorous convection, so that there is a well-mixed interior temperature  $\bar{T}(t)$  in the liquid.



Recall from earlier sections that high Rayleigh number convection has heat flux

$$F_T = \lambda k \left( \frac{\rho_0 g \alpha}{\kappa \mu} \right)^{1/3} (\bar{T} - T_m)^{4/3}$$

Consider a control volume around the solid–liquid interface *and* the thermal boundary layer, and impose energy conservation. Note that the control volume is in a frame of reference moving at the melt rate  $v$  with the interface.



Conservation of energy with melt rate  $v$  gives

$$\rho(H_l - H_s)v + \rho c_p(\bar{T} - T_m)v + F_T - k \left. \frac{\partial T}{\partial z} \right|_{a^-} = 0$$

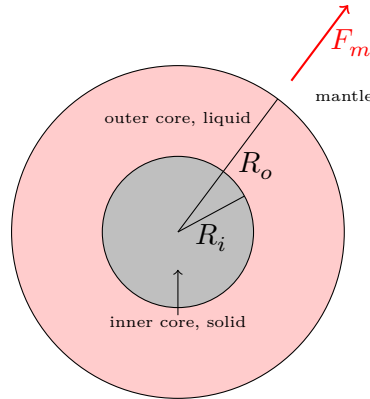
Rearranging, we have

$$\rho [L + c_p(\bar{T} - T_m)] v = k \left. \frac{\partial T}{\partial z} \right|_{a^-} - F_T$$

The first term is the release of latent heat, the second is cooling across the boundary layer, the third is conduction in the solid and the final term is the convective heat flux. This is the modified Stefan condition (needed for ES1 Q4).

#### 4.5 Earth's core

Consider solidification of the Earth's inner core, driven by heat flux  $F_m$  from the mantle.



We require a relation between the melt temperature  $T_m = T_m(p)$  and the pressure  $p$ , which is provided by the *Clausius-Clapeyron equation*:

$$\rho_s \frac{L}{T_m} \frac{\partial T_m}{\partial p} = \frac{\rho_s}{\rho_l} - 1$$

The RHS quantity is positive for iron in the inner core. Given a temperature profile  $T_m$  and a fixed boundary temperature, this equation gives the radius at which that boundary temperature is met, given a pressure profile  $p(r)$ . To leading order, the pressure is hydrostatic

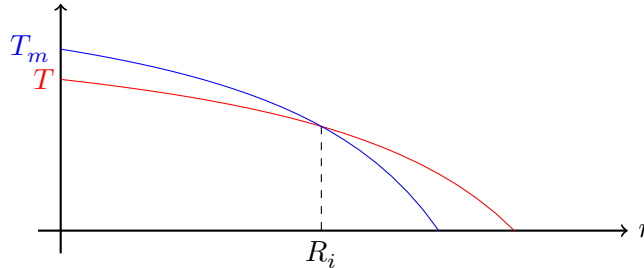
$$\begin{aligned} \frac{\partial p}{\partial r} &= -\rho g(r) = -\rho \frac{G}{r^2} \left( \frac{4\pi}{3} r^3 \rho \right) = -\frac{4\pi}{3} G \rho^2 r \\ \Rightarrow p &= p_0 - \frac{2\pi}{3} G \rho^2 r^2 \end{aligned}$$



Substituting this form of the pressure into the Clausius-Clapeyron equation gives

$$T_m \approx T_0 - \frac{2\pi GT_0(\rho_s - \rho_l)}{3L} r^2$$

The radius at which the (adiabatic) core temperature  $T$  is equal to  $T_m$  is where the solid-liquid interface exists, i.e. at  $r = R_i$ .



Note we separately assume a well mixed adiabatic temperature profile in the outer core  $\bar{T} = T_m$ . Consider the two choices of limiting cases:

1. No heat flux from inner core
2. Perfectly conducting inner core  $T_{IC} = T_m$ . Outer core is convecting vigorously so assume well mixed,  $T$  is adiabatic, i.e. 'potential temperature' is isothermal.

The balance of heat fluxes is

$$4\pi R_0^2 F_m = \frac{4\pi}{3} [R_0^3 - R_i^3] \rho c_p \frac{\partial T_m}{\partial t} + \rho L 4\pi R_i^2 \frac{dR_i}{dt}$$

In case 2 the  $R_i^3$  term in the first RHS term is neglected. In this case,

$$4\pi R_0^2 F_m = \frac{4\pi}{3} R_0^3 \rho c_p \cdot \frac{4\pi}{3} \frac{GT_m \delta \rho}{L} R_i \dot{R}_i + \rho L 4\pi R_i^2 \dot{R}_i$$

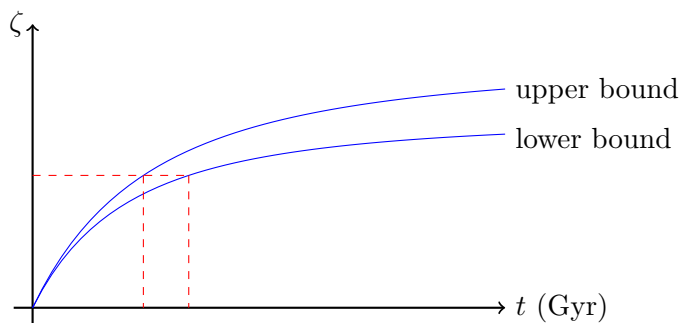
Hence the time averaged mantle heat flux is

$$\frac{1}{m} \int_0^t F_m(t') dt' = \zeta^2 + \mathcal{S} \zeta^3$$

where  $\zeta = R_i/R_o$ , the Stefan number is  $\mathcal{S} = \rho L R_o / m$  and

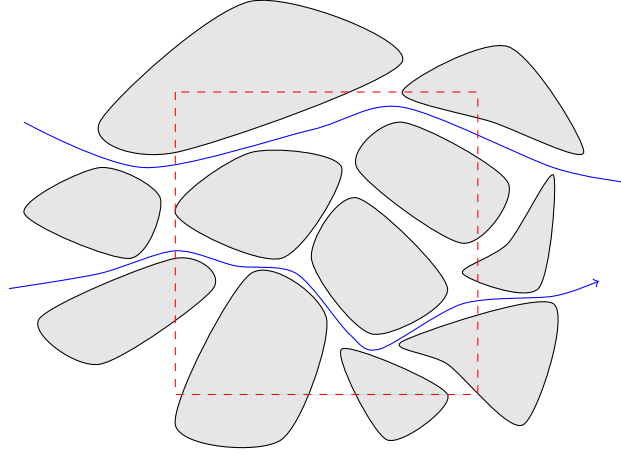
$$m = \frac{2}{9} R_0^3 \rho c_p \frac{GT_m \delta \rho}{L}$$

Real-world measurements give  $\zeta = 1/3$ ,  $\mathcal{S} = 0.4$ , so growth is controlled by the mantle heat flux balancing cooling (rather than latent heat). Note that knowing  $\zeta = 1/3$  does not allow a good estimate of the age of the inner core, given the upper and lower bounds:



## 5 Flows in porous media

To avoid the complexity of modelling very fine details, we model the flow averaged over many *pore scales* and assume viscous dissipation balances the driving forces.



Define the *porosity*

$$\phi = \frac{\text{fluid volume}}{\text{total volume}}$$

so that tightly packed media has low porosity and highly saturated media has high porosity. For context, random close-packed spheres have  $\phi \approx 0.37$  whilst hexagonally close-packed monodisperse spheres have  $\phi = 1 - \pi/3\sqrt{2} \approx 0.26$ .

### 5.1 Darcy's law

Darcy's law represents the assumption that viscous dissipation at the pore scale balances the driving forces, pressure and buoyancy. We have

$$\mu \mathbf{u} = -k(\phi)(\nabla p + \rho g \hat{\mathbf{z}})$$

where  $\mathbf{u} = \phi \tilde{\mathbf{u}}$  is the fluid flux per unit area and  $\tilde{\mathbf{u}}$  is the *interstitial fluid velocity*, i.e. the actual fluid velocity. The tensor  $k(\phi)$  is the *permeability*, which characterises the resistance to flow. In general, this is a tensor and depends on local properties; for example long grain-like particles favour motion along their long axis.  $k$  reflects the geometry/tortuosity and porosity of the medium.

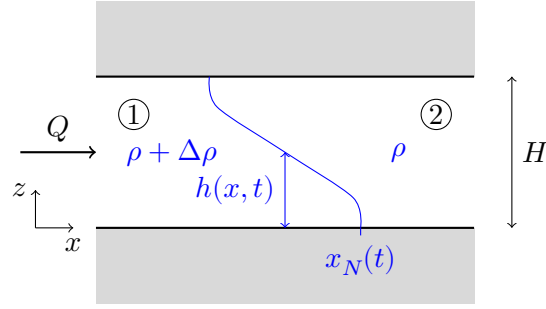
**Example.** The Kozeny-Carman equation for monodisperse random close packed spheres gives

$$k = \frac{d^2}{180} \frac{\phi^3}{(1 - \phi)^2}$$

where  $d$  is the grain size.

### 5.2 Porous gravity currents

Consider flows in a confined aquifer: we assume two fluids with density  $\rho + \Delta\rho$  and  $\rho$  separated by an interface  $z = h(x, t)$  with the nose of the gravity current at  $x = x_N(t)$  in an aquifer of height  $H$ . There is a volume flux  $Q$  injected into the aquifer and background pressure  $p_0(x, z, t)$ .



The large aspect ratio implies the flow is mainly horizontal, hence pressure is hydrostatic to leading order. The flow is driven by a combination of buoyancy ( $\sim \Delta\rho g$ ) and injection ( $\sim Q$ ), and is primarily horizontal. The pressure in the injected (lower) fluid is

$$p_1 = p_0 - (\rho + \Delta\rho)gz$$

whilst the pressure in the ambient (upper) fluid) is

$$p_2 = p_0 - (\rho + \Delta\rho)gh - \rho g(z - h)$$

The fluid flow is given by Darcy's law. For the injected fluid,

$$u_1 = -\frac{k}{\mu} \frac{\partial p_1}{\partial x} = -\frac{k}{\mu} \frac{\partial p_0}{\partial x}$$

Similarly, for the ambient fluid

$$u_2 = -\frac{k}{\mu} \frac{\partial p_2}{\partial x} = -\frac{k}{\mu} \left( \frac{\partial p_0}{\partial x} - \Delta\rho g \frac{\partial h}{\partial x} \right)$$

Now global conservation of the fluid flux gives

$$Q = u_1 h + u_2 (H - h) = -\frac{k}{\mu} \left[ \frac{\partial p_0}{\partial x} h + \frac{\partial p_0}{\partial x} (H - h) - \Delta\rho g \frac{\partial h}{\partial x} (H - h) \right]$$

Rearranging for the background pressure gradient gives the injected fluid velocity as

$$u_1 = -\frac{k}{\mu} \frac{\partial p_0}{\partial x} = \frac{Q}{H} - \frac{k\Delta\rho g}{\mu} \frac{H - h}{H} \frac{\partial h}{\partial x}$$

Local mass conservation in the lower (injected) layer gives

$$\begin{aligned} \phi \frac{\partial h}{\partial t} + \frac{\partial}{\partial x} (u_1 h) &= 0 \\ \Rightarrow \phi \frac{\partial h}{\partial t} + \frac{\partial}{\partial x} \left( \frac{Qh}{H} \right) &= \frac{k\Delta\rho g}{\mu} \frac{\partial}{\partial x} \left[ \frac{H - h}{H} h \frac{\partial h}{\partial x} \right] \end{aligned} \quad (11)$$

The first term is the change in fluid mass, the second the injection pressure driven flux, and the RHS term is the buoyancy driven flux which is associated with slumping due to the hydrostatic pressure. Note that typically the mass flux  $u_1 h$  in a viscous gravity current scales as  $\sim h^3$ , here the scaling is  $\sim h$  as the viscous drag is provided only at the bottom boundary.

Consider a moving reference frame with coordinate

$$y = x - \frac{Qt}{\phi H}$$

Thus we treat  $h$  as a function of  $y$  and  $t$ :  $h(x, t) = h(y(x, t), t)$ . Then

$$\phi \frac{\partial h}{\partial t} = u_b \frac{\partial}{\partial y} \left[ \frac{H-h}{H} h \frac{\partial h}{\partial y} \right]$$

where  $u_b = k\Delta\rho g/\mu$  is the *buoyancy velocity*. Assuming  $h \sim H$ , scaling analysis gives

$$\phi \frac{H}{t} \sim u_b \frac{H^2}{y^2} \implies y \sim \left( \frac{u_b H t}{\phi} \right)^{1/2}$$

Hence introduce similarity variable  $\zeta$  defined by

$$\zeta = \frac{y}{(u_b H t / \phi)^{1/2}}$$

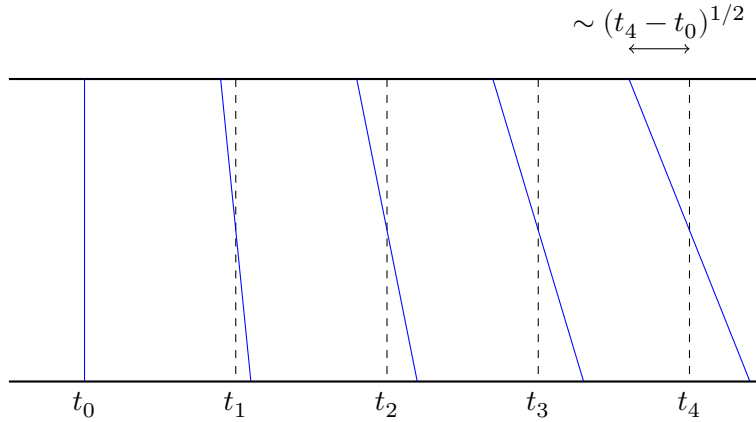
and similarity solution  $h = Hf(\zeta)$  where  $f$  satisfies

$$-\frac{\zeta}{2} f' = [(1-f)ff']' \quad (12)$$

with  $-\lambda < \zeta < \lambda$  for some eigenvalue  $\lambda$  to be determined. The solution is

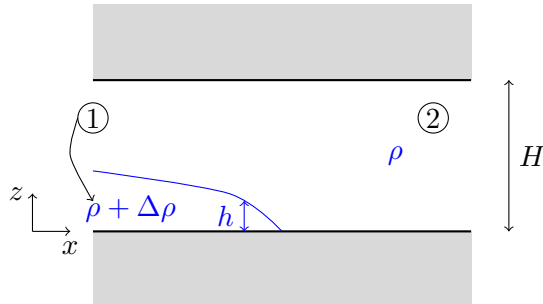
$$f = \frac{1}{2} \left( 1 - \frac{\zeta}{\lambda} \right)$$

which upon substitution into (12) gives  $\lambda = 1$ . The interface slumps with width  $\sim t^{1/2}$  in the travelling frame of reference.



### 5.3 Constant volume gravity current (porous & unconfined)

Consider the limit of the above problem  $h \ll H$ . This applies at early times during injection or late times during constant volume expansion.



Recall the injection flux  $Q$  can be written as

$$Q = -\frac{k}{\mu} \left[ H \frac{\partial p_0}{\partial x} - \Delta \rho g (H - h) \frac{\partial h}{\partial x} \right]$$

In the unconfined limit we have  $h \ll H$  and  $Q/H \approx 0$ . Thus

$$\frac{\partial p_0}{\partial x} = \Delta \rho g \frac{\partial h}{\partial x}$$

Hence the apparent injected fluid velocity in region 1 is

$$u_1 \approx -\frac{k}{\mu} \frac{\partial p_0}{\partial x} = -\frac{k \Delta \rho g}{\mu} \frac{\partial h}{\partial x}$$

The governing equation for  $h$  following from (11) is

$$\phi \frac{\partial h}{\partial t} = \frac{k \Delta \rho g}{\mu} \frac{\partial}{\partial x} \left( h \frac{\partial h}{\partial x} \right)$$

and the flux in the gravity current is

$$q = \int_0^h u_1 \, dz = -\frac{k \Delta \rho g}{\mu} h \frac{\partial h}{\partial x}$$

Boundary conditions at the nose are  $h(x_N) = 0$  and

$$q(x_N) = 0 = -u_b h \left. \frac{\partial h}{\partial x} \right|_{x_N}$$

Despite the 2nd order governing equation, the undetermined nose position  $x_N$  means a third condition is required. Global mass conservation is expressed as

$$V = \phi \int_0^{x_N} h \, dx$$

where  $V$  is the prescribed volume per unit width. This is equivalent to a condition on the flux at  $x = 0$ . The scalings are

$$\begin{aligned} \phi \frac{h}{t} &\sim u_b \frac{h^2}{x^2}, & V &\sim \phi h x \\ \Rightarrow x &\sim \left( \frac{u_b V}{\phi^2} \right)^{1/3} t^{1/3} \end{aligned}$$

Hence we define a similarity variable:

$$\eta = \frac{x t^{-1/3}}{(u_b V / \phi^2)^{1/3}}, \quad h = \left( \frac{V^2}{u_b \phi} \right)^{1/3} t^{-1/3} f(\eta)$$

where  $0 \leq \eta \leq \eta_N$ . The governing equation for  $f(\eta)$  is then

$$\begin{aligned} -\frac{1}{3} f - \frac{\eta}{3} f' &= (f f')' \\ \Rightarrow \eta f &= -3 f f' + c \end{aligned}$$

Now  $h(x_N) = 0$  implies  $c = 0$ . Hence  $f' = -\eta/3$ . The condition  $q(x_N) = 0$  gives

$$f = \frac{1}{6} (\eta_N^2 - \eta^2)$$

Global conservation of mass requires

$$1 = \int_0^{\eta_N} f d\eta = \frac{1}{6} \left( \eta_N^3 - \frac{1}{3} \eta_N^3 \right) = \frac{1}{9} \eta_N^3$$

Thus the nose is located at  $\eta_N = 9^{1/3} \approx 2.08$ , or in terms of the original coordinates

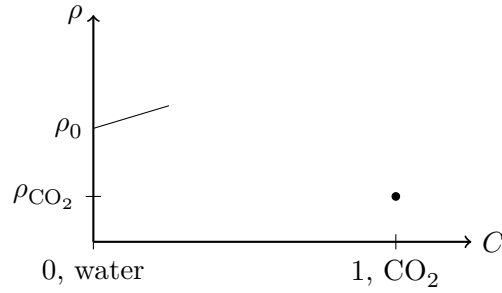
$$x_N = 2.08 \left( \frac{u_b V t}{\phi^2} \right)^{1/3}$$

## 6 Convection in porous media

A common example of convection in porous media is geothermal dissolution of  $CO_2$ . The sequestration of  $CO_2$  into water changes the density to

$$\rho = \rho_0(1 + \beta C)$$

where  $C$  is the concentration of  $CO_2$ .



The natural buoyancy velocity is then

$$u_b \equiv \frac{\Delta \rho g k}{\mu} = \frac{\rho_0 \beta \Delta C g k}{\mu}$$

Convection is driven by dissolution of  $CO_2$  into water. The timescale to convect over a distance  $h$  is

$$\tau_{\text{ad}} \sim \frac{h}{u_b}$$

whilst the timescale to diffuse over a distance  $h$  is

$$\tau_{\text{diff}} \sim \frac{h^2}{D}$$

where  $D$  is the diffusivity. We have convection if  $\tau_{\text{ad}} < \tau_{\text{diff}}$ , hence

$$\frac{h}{u_b} < \frac{h^2}{D} \implies \text{Ra} = \frac{u_b h}{D} = \frac{\Delta \rho g k h}{\mu D}$$

Note: it can be shown via similar analysis to pure convection that the onset of porous convection is at  $\text{Ra}_{\text{crit}} = 4\pi^2$ .

### 6.1 Scaling for high Ra porous convection

For  $Ra \gg Ra_c$  convective flux becomes independent of the depth of domain, as the interior is statistically well mixed as with high Ra pure convection. The Nusselt number is

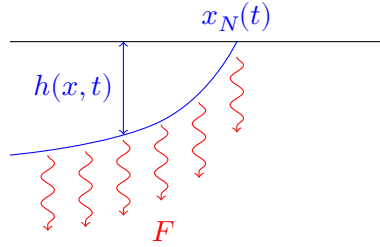
$$Nu = \frac{F}{D\Delta C/h} \propto h \sim Ra = \lambda Ra = \frac{\lambda \rho_0 g \beta \Delta c k h}{D\mu}$$

where  $F$  is the convective flux. We therefore have

$$F = \frac{\lambda \rho_0 g \beta k}{\mu} \Delta C^2 = \lambda u_b \Delta C$$

### 6.2 Dissolution from an unconfined aquifer

Consider a convective flux from a constant volume gravity current in an unconfined aquifer. Note that by constant volume we mean no input flux; the volume will decrease due to convection.



Since the problem is set in an unconfined aquifer we may assume the convective flux  $F$  is constant. The governing equation is then

$$\phi \frac{\partial h}{\partial t} - u_b \frac{\partial}{\partial x} \left( h \frac{\partial h}{\partial x} \right) = -F$$

Denote the initial volume of the gravity current as  $V$  (note this is a 2D volume). Non-dimensionalise the equations with scalings  $h, x \sim \sqrt{V/\phi}$  and  $t \sim u_b^{-1} \sqrt{V/\phi}$ .

$$h_t - (hh_x)_x = -\varepsilon$$

where  $\varepsilon = F/u_b$ . Recall that for  $\varepsilon = 0$  we get a self-similar solution:

$$h_{\varepsilon=0} = t^{-1/3} f(xt^{-1/3}) = t^{-1/3} \cdot \frac{1}{6}(\eta_N^2 - \eta^2)$$

where  $\eta = xt^{-1/3}$  and  $\eta_N = 9^{1/3}$ . For this problem we try ansatz  $h = h_0 + ct$  for some undetermined constant  $c$ . Then

$$\begin{aligned} (h_0 + ct)_t &= ((h_0 + ct)(h_0)_x)_x - \varepsilon \\ \Rightarrow \cancel{h_0}_t + c &= \cancel{(h_0 h_{0x})}_x + cth_{0xx} - \varepsilon \end{aligned}$$

Thus  $c$  satisfies

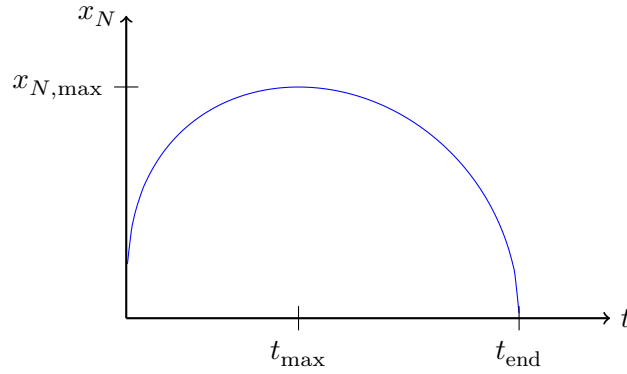
$$c = ct \cdot \frac{t^{-1/3}}{t^{2/3}} \left( -\frac{2}{6} \right) - \varepsilon \Rightarrow c = -\frac{3\varepsilon}{4}$$

The full solution is

$$h = \frac{1}{6}t^{-1/3} \left( 9^{2/3} - \frac{x^2}{t^{2/3}} \right) - \frac{3\varepsilon t}{4}$$

The extent of the current  $x_N(t)$  is given by  $h = 0$

$$\Rightarrow x_N(t) = \left( 9^{2/3}t^{2/3} - \frac{9\varepsilon}{2}t^2 \right)^{1/2}$$



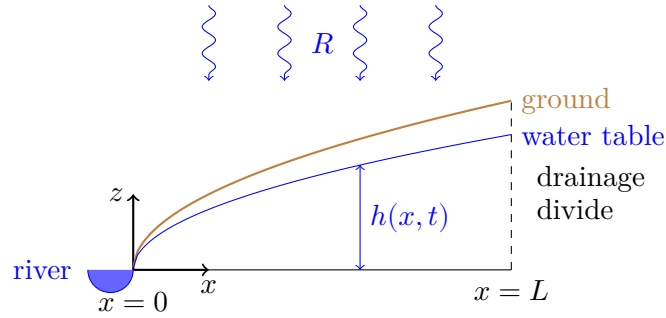
The maximum extent and the time at which it is reached are

$$x_{N,\max} = \left( \frac{2^3}{3\varepsilon} \right)^{1/4}$$

$$t_{\text{end}} = \left( \frac{2}{\varepsilon} \right)^{3/4} 3^{-1/2}$$

### 6.3 Charging/discharging of a groundwater aquifer

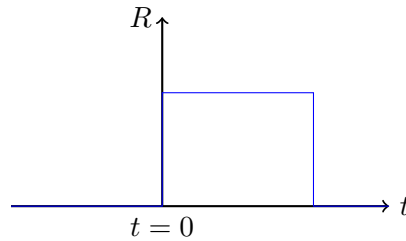
Consider rainfall at rate  $R$  charging a groundwater aquifer with depth  $h(x, t)$ .



The groundwater table evolution is given by

$$\phi \frac{\partial h}{\partial t} = u_b \frac{\partial}{\partial x} \left( h \frac{\partial h}{\partial x} \right) + R$$

where  $R$  is a top-hat function representing a period of rainfall:



At the river, we assume  $h(0, t) = 0$ . At the drainage divide, which is the boundary between catchment basins,  $h_x|_{x=L} = 0$  by definition.



**Rainfall ( $R \neq 0$ ).** At early times when rain is falling, we start from  $h(x, 0) \approx 0$ . The far-field water table is  $h_\infty \approx Rt/\phi$ . Note at early times, the extent of the water table is not important. Near the river,

$$\frac{\phi h}{t} \sim u_b \frac{h^2}{x^2} \sim R$$

Thus  $h$  and  $x$  scale as

$$h \sim \frac{Rt}{\phi}, \quad x \sim \left(\frac{u_b}{R}\right)^{1/2} h \sim \left(\frac{u_b}{R}\right)^{1/2} \frac{Rt}{\phi} \sim \left(\frac{u_b R}{\phi^2}\right)^{1/2} t$$

The flux into the river is

$$Q(0) = u_b h \left. \frac{\partial h}{\partial x} \right|_{x=0} \sim u_b \frac{h^2}{x} \sim \frac{u_b^{1/2} R^{3/2}}{\phi} t$$

Note: there is a self-similar solution for the evolution near the river at early times, with similarity variable  $\eta$  and depth  $h$  defined by

$$\eta = \frac{x}{(u_b R/\phi^2)^{1/2} t}, \quad h = \frac{Rt}{\phi} f(\eta)$$

where  $f(\eta)$  satisfies

$$f - \eta f' = (ff')' + 1$$

subject to  $f(0) = 0, f'(\eta \rightarrow \infty) = 0$ . The term ‘early times’ here refers to  $x \ll L$ , or equivalently

$$t \ll \frac{\phi L}{(u_b R)^{1/2}}$$

At late times, after a sustained period of rainfall, the solution reaches a steady state  $\partial_t = 0$  with

$$\frac{h}{L} = \left(\frac{R}{u_b}\right)^{1/2} \sqrt{\frac{2x}{L} - \left(\frac{x}{L}\right)^2}$$

The flux into the river in this steady state regime is

$$Q = RL$$

**Aquifer discharge ( $R = 0$ )** When  $R = 0$ , assuming  $x \sim L$  (as there is no distinction close or far from the river without rainfall) we have scalings

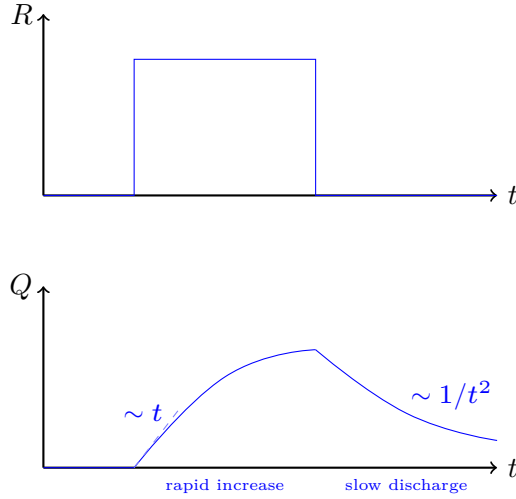
$$\frac{\phi h}{t} \sim u_b \frac{h^2}{L^2} \implies h \sim \frac{\phi L^2}{u_b t}$$

Use similarity solution  $h = \frac{\phi L^2}{u_b t} g\left(\frac{x}{L}\right)$  where  $g$  satisfies

$$-g = (gg')'$$

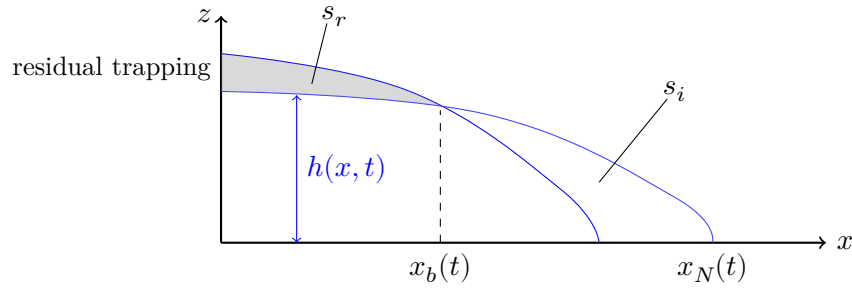
subject to  $g(0) = 0$  (zero height at the river) and  $g'(1) = 0$  (no flux across drainage divide). The flux into the river is

$$Q = u_b h \left. \frac{\partial h}{\partial x} \right|_{x=0} \sim \frac{\phi^2 L^3}{u_b} t^{-2}$$



### 6.4 Gravity currents with residual trapping

Consider surface tension acting in a groundwater aquifer such that part of the groundwater amount is *residually trapped* in the pores that the gravity current migrates through.



Define saturation by

$$s = \frac{\text{pore space filled with water}}{\text{total pore space}}$$

We will use two saturations in this problem: the initial saturation  $s_i$  where the gravity current first reaches a region of the porous medium, and a residual saturation  $s_r$  after the current has passed through.

The governing equations for the *mobile* fluid are

$$\begin{aligned} \phi s_i h_t &= u_b (h h_x)_x & x_b < x < x_N, \quad h_t > 0 \\ \phi s_i h_t &= u_b (h h_x)_x + \phi s_r h_t & 0 < x < x_b, \quad h_t < 0 \end{aligned}$$

Note that in the region  $0 < x < x_b$  the height is decreasing so  $\phi s_r h_t$  is a *loss* term. The buoyancy velocity  $u_b = \Delta \rho g k / \mu$  as usual for a gravity current. The boundary conditions on  $h$  are

$$\begin{aligned} h_x(0) &= 0 & \text{no input flux} \\ h(x_N) &= 0 & \text{zero height at nose} \\ -u_b h h_x|_{x_N} &= 0 & \text{zero mass flux through nose} \\ [h(x_b)]_-^+ &= 0 & \text{continuity of height} \\ [-u_b h h_x|_{x_b}]_-^+ &= 0 & \text{continuity of mass flux} \\ h_t(x_b) &= 0 & \text{defines boundary } x_b \end{aligned}$$

We also supplement the problem with a constant volume initial condition

$$V = \phi s_i \int_0^{x_N(0)} h(x, 0) dx$$

To simplify notation, define

$$\mathcal{U}_i = \begin{cases} \mathcal{U}_1 = \frac{u_b}{\phi s_i} & h_t > 0 \\ \mathcal{U}_2 = \frac{\phi s_i u_b}{\phi(s_i - s_r)} & h_t < 0 \end{cases}$$

$$\mathcal{V} = \frac{V}{\phi s_i} = \int_0^{x_N(0)} h(x, 0) dx$$

so that the governing equation is

$$h_t = \mathcal{U}_i (h h_x)_x$$

subject to the same boundary conditions, except  $\mathcal{U}_1$  replaces  $u_b$  in the mass flux conditions. Recall that for  $s_r = 0$ ,  $\mathcal{U}_1 = \mathcal{U}_2$  with solution

$$h = \left( \frac{\mathcal{V}^2}{\mathcal{U}_1 t} \right)^{1/3} f(\eta), \quad \eta = \frac{x}{(\mathcal{U}_1 \mathcal{V} t)^{1/2}}, \quad f = \frac{1}{6} (9^{2/3} - \eta^2)$$

For  $s_r \neq 0$ ,  $\mathcal{U}_1 \neq \mathcal{U}_2$  and the initial condition is remembered, so we look for an incomplete self-similar solution, known as a self-similar solution of the *second type*. Define similarity variable

$$\eta_0 = \frac{x_0}{(\mathcal{U}_1 \mathcal{V} t)^{1/3}}$$

where  $x_0 = x_N(0)$  and specify  $h$  with an undetermined parameter  $\gamma$  as

$$h = \eta_0^\gamma \left( \frac{\mathcal{V}^2}{\mathcal{U}_1 t} \right)^{1/3} f\left(\frac{\eta}{\eta_0^\gamma}, \frac{\mathcal{U}_1}{\mathcal{U}_2}\right)$$

To simplify notation we define  $\zeta = x/Bt^\beta$ ,  $\tilde{\mathcal{U}} = \mathcal{U}_1/\mathcal{U}_2$  and  $h = At^{-\alpha} f(\zeta, \tilde{\mathcal{U}})$  (where  $A$  and  $B$  may be determined by comparison with the original definitions) so that the governing equation gives

$$-\alpha At^{-\alpha-1} f - \frac{\beta \zeta}{t} At^{-\alpha} f' = \mathcal{U}_i \frac{A^2 t^{-2\alpha}}{B^2 t^{2\beta}} (f f')'$$

which simplifies to

$$-\frac{B^2}{A} t^{2\beta+\alpha-1} (\alpha f + \zeta \beta f') = \mathcal{U}_i (f f')'$$

Note that by introducing  $B$  and  $A$ , one of them is a free parameter, which we choose to satisfy  $B^2/A = \mathcal{U}_2$ . For self-similar behaviour we need

$$\alpha + 2\beta - 1 = 0 \implies \alpha = 1 - 2\beta$$

The governing equations are therefore

$$\begin{aligned} (f f')' + \beta \zeta f' + (1 - 2\beta) f &= 0 & h_t < 0 \\ \tilde{\mathcal{U}} (f f')' + \beta \zeta f' + (1 - 2\beta) f &= 0 & h_t > 0 \end{aligned}$$

with boundary conditions

$$\begin{aligned} f(\zeta_N) &= 0 \\ f'(0) &= 0 \\ -ff'|_{\zeta_N} &= 0 \\ [f(\zeta_b)]_-^+ &= 0 \\ [-ff'|_{\zeta_b}]_-^+ &= [f'(\zeta_b)]_-^+ = 0 \end{aligned}$$

The governing equations are satisfied only for eigenvalues  $\beta$ . The boundary position  $\zeta_b$  is then determined by

$$h_t(x_b) = 0 \implies \beta\zeta_b f'(\zeta_b) + (1 - 2\beta)f(\zeta_b) = 0$$

**Small trapping limit.** Consider the limit of small trapping  $s_r \ll s_i$ , where we can make analytical progress. Let  $U_1 = U_2(1 - \varepsilon)$  with  $\varepsilon \ll 1$ . Then

$$\tilde{U} \equiv \frac{U_1}{U_2} = 1 - \varepsilon$$

The rate of change of total volume of mobile fluid is

$$\begin{aligned} \frac{d}{dt} \int_0^{x_N} h \, dx &= \int_0^{x_N} U_i (hh_x)_x \, dx \\ &= U_2 hh_x|_{x_b^-} - U_1 hh_x|_{x_b^+} \end{aligned} \quad (13)$$

Using the form  $h = At^{-\alpha} f(\zeta, 1 - \varepsilon)$ , with  $\zeta = x/Bt^\beta$ , gives

$$\begin{aligned} \frac{d}{dt} \int_0^{x_N} h \, dx &= \frac{d}{dt} \left[ ABt^{\beta-\alpha} \int_0^{x_N} f \, d\zeta \right] \\ &= (\beta - \alpha) ABt^{\beta-\alpha-1} \int_0^{\zeta_N} f \, d\zeta \\ &= U_2 \frac{(At^{-\alpha})^2}{Bt^\beta} \left[ ff'|_{\zeta_b^-} - (1 - \varepsilon) ff'|_{\zeta_b^+} \right] \end{aligned}$$

where the final equality follows from (13). Now using  $B^2/A \equiv U_2$ ,  $\alpha = 1 - 2\beta$  and the boundary condition  $ff'$  continuous at  $\zeta = \zeta_b$ , we have

$$(3\beta - 1) \int_0^{\zeta_N} f \, d\zeta = \varepsilon ff'|_{\zeta_b^+} \quad (14)$$

When  $\varepsilon = 0$ , we have  $\beta = 1/3$  exactly. If  $\varepsilon = 0$  we also have  $f = (\zeta_N^2 - \zeta^2)/6$  with  $\zeta_N = 9^{1/3}$ . The condition  $h_t(x_b) = 0$  gives

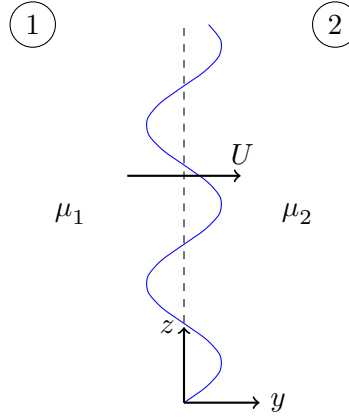
$$\begin{aligned} f|_{\zeta_b} + \zeta_b f'|_{\zeta_b} &= 0 \\ \implies \zeta_b &= \zeta_N/\sqrt{3} \end{aligned}$$

Hence assuming  $\beta$  is an  $\mathcal{O}(\varepsilon)$  correction to  $1/3$ , (14) becomes

$$\begin{aligned} (3\beta - 1) \int_0^{\zeta_N} \frac{1}{6} (\zeta_N^2 - \zeta^2) d\zeta &= \varepsilon \cdot \frac{1}{6} (\zeta_N^2 - \zeta_b^2) \cdot -\frac{2}{6} \zeta_b \\ \Rightarrow \beta &= \frac{1}{3} - \frac{\varepsilon}{3^{5/2}} \\ &= \frac{1}{3} - \frac{s_r/s_i}{3^{5/2}} \end{aligned}$$

## 7 Saffman-Taylor fingering

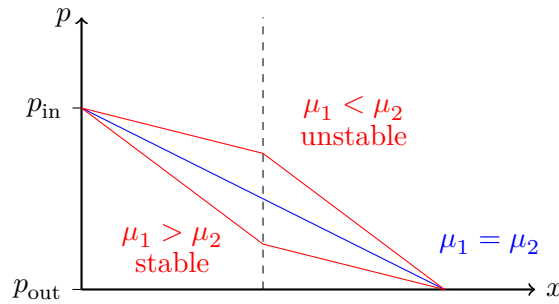
*Saffman-Taylor fingering* (also *viscous fingering*) is an instability between fluid of viscosity  $\mu_1$  injected into a porous medium saturated with fluid of viscosity  $\mu_2$ . In the case of a solid-liquid interface, this is also called *Mullins-Sekerka fingering* (or *morphological instability*). .



**Basis for instability.** Darcy's law gives  $\mathbf{u}_i = -k/\mu_i \nabla p_i$  for  $i = 1, 2$ . Since the interface moves at velocity  $U$  we have

$$U = u_1 = -\frac{k}{\mu_1} \frac{\partial p_1}{\partial x} = u_2 = -\frac{k}{\mu_2} \frac{\partial p_2}{\partial x}$$

If  $\mu_1 < \mu_2$  then the pressure gradient is greatest in fluid 2, so the fingering perturbations grow and we have instability. If instead  $\mu_1 > \mu_2$  then the perturbations decay and we have stability.



**Pressure fields.** Consider a travelling frame of reference  $y = x - \frac{Ut}{\phi}$ . Then

$$\begin{aligned} p_1 &= -\frac{\mu_1 U}{k} y + p_0 \\ p_2 &= -\frac{\mu_2 U}{k} y + p_0 \end{aligned}$$

Now consider perturbations to the interface of the form

$$y = \eta(z, t) = \tilde{\eta} e^{i\alpha z + \sigma t}$$

Darcy's law and incompressibility (equivalently conservation of mass) give

$$\mathbf{u}_i = -\frac{k}{\mu_i} \nabla p_i, \quad \nabla \cdot \mathbf{u}_i = 0 \implies \nabla^2 p_i = 0$$

Hence the pressure field, including perturbations, must be of the form

$$p_1 = p_0 - \frac{\mu_1 U}{k} y + \hat{p}_1(y) e^{i\alpha z + \sigma t}$$

Requiring  $\nabla^2 p_1 = 0$  gives

$$\begin{aligned} \frac{\partial^2 \hat{p}_1}{\partial y^2} - \alpha^2 \hat{p}_1 &= 0 \\ \implies \hat{p}_1(y) &= A e^{\alpha y} \end{aligned}$$

where we have used the requirement that perturbations decay away from the interface. Applying the same theory to  $p_2$  gives the pressure fields as

$$\begin{aligned} p_1 &= p_0 - \frac{\mu_1 U}{k} y + A e^{i\alpha z + \alpha y + \sigma t} \\ p_2 &= p_0 - \frac{\mu_2 U}{k} y + A e^{i\alpha z - \alpha y + \sigma t} \end{aligned}$$

Next, match pressures and velocities at the interface. Pressure continuity along the perturbed interface requires

$$\begin{aligned} p_1(\eta) &= p_2(\eta) \\ \implies p_1|_0 + \frac{\partial p_1}{\partial y} \Big|_0 \hat{\eta} &= p_2|_0 + \frac{\partial p_2}{\partial y} \Big|_0 \hat{\eta} \\ \implies p_0 + A + \left( -\frac{\mu_1 U}{k} + \alpha A \right) \hat{\eta} &= p_0 + B + \left( -\frac{\mu_2 U}{k} - \alpha B \right) \hat{\eta} \end{aligned}$$

Cancelled terms are negligible since  $A, B = \mathcal{O}(\hat{\eta})$ , i.e. terms are second order. Velocity continuity at the interface requires

$$[\mathbf{u} \cdot \hat{\mathbf{n}}]_{y=\eta} = 0$$

where  $\hat{\mathbf{n}}$  is the unit normal to the interface. Taylor expanding as before, we have

$$u_1|_0 + \frac{\partial u_1}{\partial y} \Big|_0 \hat{\eta} = u_2|_0 + \frac{\partial u_2}{\partial y} \Big|_0 \hat{\eta}$$

Now note

$$\phi \frac{\partial \eta}{\partial t} = \phi \sigma \hat{\eta} = -\frac{k}{\mu_1} \alpha A = \frac{k}{\mu_2} \alpha B$$

Hence the velocity continuity condition becomes

$$\begin{aligned} -\frac{\mu_1 \phi \sigma}{k \alpha} - \frac{\mu_1 U}{k} \hat{\eta} &= \frac{\mu_2 \phi \sigma}{k \alpha} \hat{\eta} - \frac{\mu_2 U}{k} \hat{\eta} \\ \implies \sigma &= \frac{\alpha U}{\phi} \frac{\mu_2 - \mu_1}{\mu_2 + \mu_1} \end{aligned}$$

Hence the system is unstable for  $\mu_2 > \mu_1$ , as expected. Most unstable for  $\alpha \rightarrow \infty$ , i.e. the smallest wavelengths (UV catastrophe). Thus regularisation needed: additional physics at the smallest wavelengths, e.g. surface tension. Surface tension introduces a pressure discontinuity at the interface

$$p_1 - p_2 = \gamma \kappa = 2\gamma \nabla \cdot \hat{\mathbf{n}}$$

where  $\hat{\mathbf{n}}$  is the normal to the interface as before. Hence

$$\hat{\mathbf{n}} = \frac{\nabla(y - \eta)}{|\nabla(y - \eta)|} \approx \hat{\mathbf{y}} - \eta_z \hat{\mathbf{z}}$$

Then  $\nabla \cdot \hat{\mathbf{n}} = -\eta_{zz}$ , so the pressure condition becomes

$$\begin{aligned} p_1|_0 + \left. \frac{\partial p_1}{\partial y} \right|_0 \hat{\eta} &= p_2|_0 + \left. \frac{\partial p_2}{\partial y} \right|_0 \hat{\eta} - 2\gamma \eta_{zz} \\ \Rightarrow p_0 + A - \frac{\mu_1 U}{k} \hat{\eta} &= p_0 + B - \frac{\mu_2 U}{k} \hat{\eta} + 2\gamma \alpha^2 \hat{\eta} \end{aligned}$$

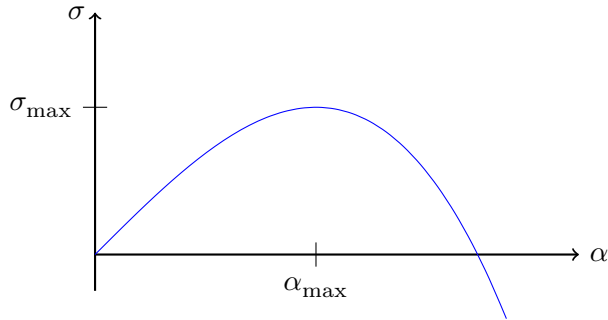
where the second order terms are dropped as before. The growth rate is then

$$\begin{aligned} \sigma &= \frac{\alpha U}{\phi} \left( \frac{\mu_2 - \mu_1}{\mu_2 + \mu_1} - \alpha^2 \frac{2\gamma k}{(\mu_2 + \mu_1)U} \right) \\ &= \frac{\alpha U}{\phi} \left( m - \frac{m+1}{\text{Ca}} k \alpha^2 \right) \end{aligned}$$

where  $m$  is the *mobility ratio*

$$m = \frac{\mu_2 - \mu_1}{\mu_2 + \mu_1}$$

and  $\text{Ca} = \mu_2 U / \gamma$  is the *capillary number*.



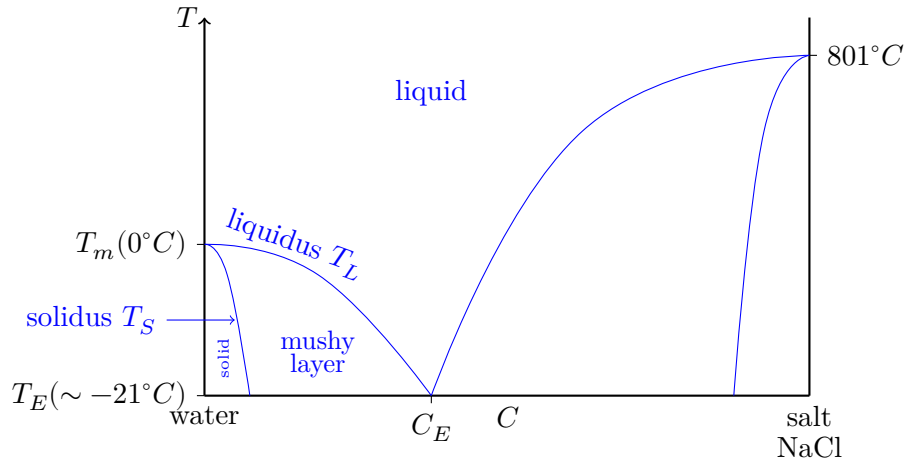
Maximum growth rate occurs at wavelength

$$\alpha_{\max} = \left( \frac{\text{Ca } m}{3(m+1)} \right)^{1/2} k^{-1/2}$$

Surface tension suppresses the smallest wavelengths, providing wavelength selection for the instability.

## 8 Solidification of binary eutectic mixtures

A natural example is the solidification of sea ice in polar oceans. The problem is an extension of previous solidification problems with concentration as an additional species to handle. Consider a phase diagram of temperature and concentration:



- The freezing/melting temperature or *liquidus* is a function of concentration:

$$T_L = T_L(C)$$

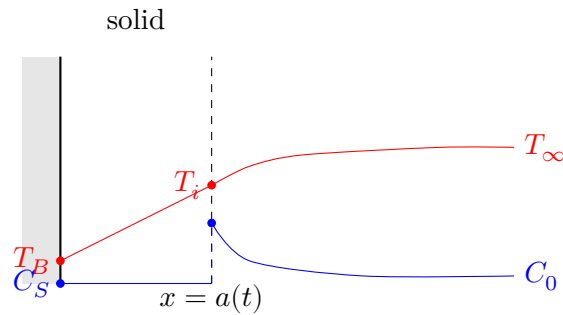
- The solution is solid below the *solidus*

$$T_S(C) = k_D T_L(C)$$

where  $k_D$  is the *partition coefficient*. For ice,  $k_D \approx 0$  so ice is ‘pure’ water.

- Between the solidus and liquidus, solid and liquid phases coexist in equilibrium.
- The *eutectic temperature*  $T_E$  is the temperature at which both water and salt form ice crystals. Below  $T_E$ , a solid solution forms.

For a freezing system, the interfacial conditions are now obtained by *both* temperature and concentration considerations.



Note  $T_i = T_L(C_i)$  and take  $C_S = 0$  i.e. the solid is pure ice. Assume a liquidus of the form

$$T_L(C) = -mC$$

so that  $T_m = 0^\circ\text{C}$ . Temperature is governed by

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial x^2} \quad 0 < x < \infty$$



where we have assumed equal thermal properties in each phase. Concentration is governed by

$$\frac{\partial C}{\partial t} = D \frac{\partial^2 C}{\partial x^2} \quad a < x < \infty$$

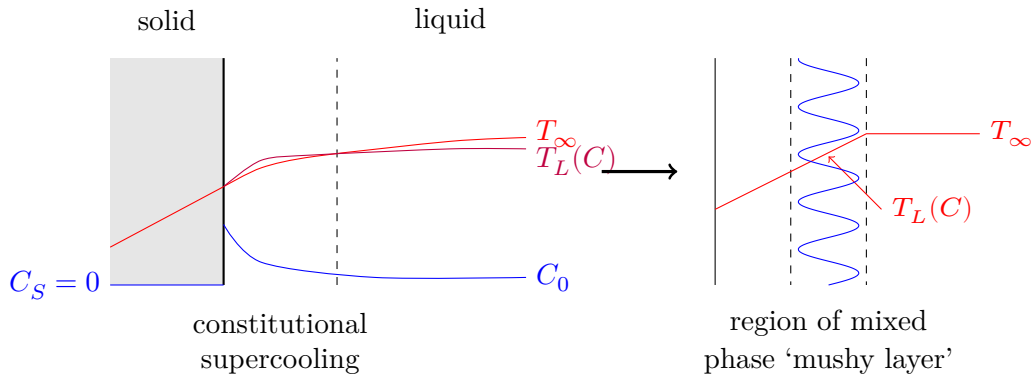
where  $D$  is the salt diffusivity. The interfacial conditions are

$$\begin{aligned} T_i &= T_L(C_i) = -mC_i \\ \rho L \dot{a} &= k \left. \frac{\partial T}{\partial x} \right|_{a^-} - k \left. \frac{\partial T}{\partial x} \right|_{a^+} && \text{Stefan condition} \\ (C_i - C_S) \dot{a} &= -D \left. \frac{\partial C}{\partial x} \right|_{a^+} && \text{Conservation of solute} \end{aligned}$$

### 8.1 Morphological instability

Consider the case of solidification & viscous fingering at a solid-liquid interface, known as morphological instability. This instability:

- increases surface area of the interface
- enhances the release of latent heat
- raises temperature  $T$
- enhances rejection of solute (within interstices)

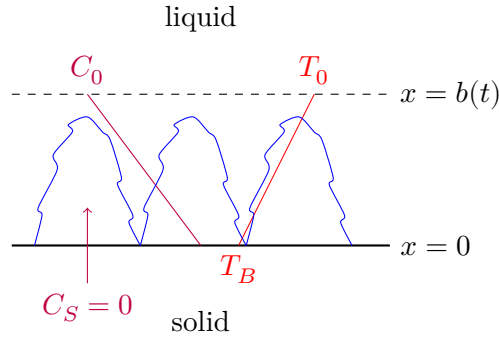


The phenomenon leads to a porous medium in thermodynamic equilibrium with a *solid fraction*  $\phi$  (note the clash of notation!) determined by local  $T, C$  via liquidus  $T_L(C)$ .

**Global conservation of solute.** Conservation of solute in the mushy layer between  $0 \leq x \leq b(t)$  is expressed by

$$\int_0^{b(t)} (1 - \phi) C_L(T) dx = C_0 b(t)$$

where  $\phi(x, t)$  is the local volume fraction of solid, i.e.  $1 - \phi$  is the *liquid fraction*. Here,  $C_L(T)$  is the liquidus concentration determined by the temperature. That is, in the mushy layer we have interstices of pure solid with  $C_S = 0$  and  $C_L$  is the concentration of the liquid in between.  $C_0$  is the concentration of the solute in the liquid.



We will use a simplified model of *bulk* properties:

1. Take quasi-static approximation for temperature:

$$T = T_B + (T_0 - T_B) \frac{x}{b}$$

where  $T_0 = T_L(C_0)$ .

2. All impurities are accommodated within the mushy layer, i.e.  $C = C_0$  for  $x > b(t)$ .
3. Linear liquid approximation

$$T_L = T_0 - m(C - C_0)$$

4. Ignore heat transfer in liquid

**Box model.** Consider the ‘trial function’  $\phi = \text{const.}$  in global solute conservation. We have

$$\int_0^b (1 - \phi) C_L(T) dx \approx (1 - \phi) \frac{C_B + C_0}{2} b = C_0 b$$

Then the solid fraction may be written

$$\phi = \frac{1}{2\mathcal{C} - 1}$$

where  $\mathcal{C} = C_B / (C_B - C_0)$ . Note when  $\mathcal{C} \gg 1$ , we can approximate  $\phi \approx 1/2\mathcal{C} \approx \mathcal{O}(\mathcal{C}^{-1})$ . With the same trial function, the Stefan condition gives

$$\phi \rho_S L \dot{b} = \bar{k}(\phi) \frac{T_0 - T_B}{b}$$

where  $\bar{k} = k_s \phi + k_l(1 - \phi)$  (which is exact for *lamella* parallel to  $\nabla T$  i.e. mushy layer composed of parallel layers). The boundary position is then

$$b = \left( \frac{2}{\mathcal{S}} \frac{\bar{k}}{k_s \phi} \kappa_s t \right)^{1/2}$$

where  $\mathcal{S}$  is the Stefan number defined by

$$\mathcal{S} = \frac{L}{c_p(T_0 - T_B)}$$

Note that for  $k_l \ll k_s$  we have

$$\frac{\bar{k}}{k_s \phi} = 1 + \frac{k_l}{k_s} \frac{1 - \phi}{\phi} \approx 1$$

For ice,  $k_s \approx 5k_l$ . Also note: growth of mushy layer scales as  $b \propto \sqrt{\kappa t}$  which is constrained by the thermal field, *not* diffusion of solute.

## 8.2 ‘Ideal’ mushy layer

Take  $\rho, c_p, k$  independent of phase. The conservation equations are then:

- Conservation of energy

$$\rho c_p \left( \frac{\partial T}{\partial t} + \mathbf{u} \cdot \nabla T \right) = k \nabla^2 T + \rho L \frac{\partial \phi}{\partial t} \quad (15)$$

where the final term represents the internal release of latent heat.

- Conservation of solute

$$(1 - \phi) \frac{\partial C}{\partial t} + \mathbf{u} \cdot \nabla C = \nabla \cdot [D(1 - \phi) \nabla C] + C \frac{\partial \phi}{\partial t} \quad (16)$$

where the diffusion term is negligible for  $D/\kappa \ll 1$ . The final term represents the internal release of solute.

- Conservation of mass

$$\nabla \cdot \mathbf{u} = 0$$

- Conservation of momentum (Darcy’s law)

$$\mathbf{u} = -\frac{\Pi}{\mu} [\nabla p + \rho \mathbf{g}]$$

where  $\Pi$  is now being used for permeability since  $k$  is the thermal conductivity. Note care must be taken with the sign of components of  $\mathbf{g}$ : a ‘check’ is that  $\mathbf{u} = 0$  should recover hydrostatic balance.

To supplement these conservation equations we have the requirement of phase equilibrium (constrained by the liquidus)

$$T = T_L(C) = T_L(C_0) - m(C - C_0) \quad (17)$$

and the equation of state

$$\rho = \rho_0 [1 - \alpha(T - T_0) + \beta(C - C_0)]$$

## 8.3 Heat capacity for a stagnant mushy layer

Consider a mushy layer at rest  $\mathbf{u} = 0$ . Conservation of solute (16) gives

$$\begin{aligned} (1 - \phi) \frac{\partial C}{\partial t} &= C \frac{\partial \phi}{\partial t} \\ \Rightarrow \frac{\partial}{\partial t} [(1 - \phi)C] &= 0 \end{aligned}$$

Hence  $(1 - \phi)C = C_0$  is constant. Then the solid fraction is

$$\phi = 1 - \frac{C_0}{C}$$

The time derivative of  $\phi$  can then be determined as

$$\frac{\partial \phi}{\partial t} = \frac{C_0}{C^2} \frac{\partial C}{\partial t} = -\frac{C_0}{mC^2} \frac{\partial T}{\partial t}$$

using (17). Substituting into (15) we have

$$\begin{aligned} \rho c_p \left[ 1 + \frac{L}{c_p} \frac{C_0}{mC^2} \right] \frac{\partial T}{\partial t} &= k \nabla^2 T \\ \Rightarrow \rho c_{p,\text{eff}} \frac{\partial T}{\partial t} &= k \nabla^2 T \end{aligned}$$

where the effective heat capacity is

$$c_{p,\text{eff}} = c_p \left[ 1 + \frac{L}{c_p} \frac{C_0}{mC^2} \right]$$

We now partially non-dimensionalise using scalings  $T \sim \Delta T = T_0 - T_B$  and  $C \sim \Delta C = C_B - C_0$ . Define the new variable

$$\theta = \frac{T - T_0}{\Delta T} = \frac{C - C_0}{\Delta C}$$

noting  $\Delta T = m\Delta C$ . We therefore have

$$\begin{aligned} \frac{\partial \theta}{\partial t} + \mathbf{u} \cdot \nabla \theta &= \kappa \nabla^2 \theta + \mathcal{S} \frac{\partial \phi}{\partial t} \\ (1 - \phi) \frac{\partial \theta}{\partial t} + \mathbf{u} \cdot \nabla \theta &= -(\mathcal{C} - \theta) \frac{\partial \phi}{\partial t} \end{aligned}$$

where we can now write  $\mathcal{S} = L/c_p \Delta T$  and  $\mathcal{C} = C_0/\Delta C$ .

For a *stagnant* mushy layer  $\mathbf{u} = 0$  we get

$$\begin{aligned} (1 - \phi) \frac{\partial \theta}{\partial t} &= -(\mathcal{C} - \theta) \frac{\partial \phi}{\partial t} \\ \Rightarrow \frac{\partial}{\partial t} [(1 - \phi)(\mathcal{C} - \theta)] &= 0 \\ (1 - \phi)(\mathcal{C} - \theta) &= \text{const.} = \mathcal{C} \end{aligned}$$

Hence we can write the solid fraction profile as

$$\phi = \frac{-\theta}{\mathcal{C} - \theta}$$

which is determined in terms of  $\theta(x, t)$ . For  $\mathcal{C} \gg 1$ ,  $\phi \ll 1$  and  $\phi_t \approx -\frac{1}{\mathcal{C}} \theta_t$ . Hence

$$\left( 1 + \frac{\mathcal{S}}{\mathcal{C}} \right) \frac{\partial \theta}{\partial t} = \kappa \nabla^2 \theta$$

The internal release of latent heat results in a modified (non-dimensional) heat capacity.

## 8.4 Convection in mushy layers

Consider a mushy layer subject to constant heat flux and constant pressure at the lower boundary  $z = h$  where  $T = T_0$  and  $T = T_B$  at  $z = 0$  with  $T_B < T_0$ . The canonical example is sea ice.

$$\begin{array}{c} \textcolor{red}{T_B} \\ \hline z = 0 \\ \\ \hline \textcolor{red}{T_0} \\ z = h \end{array}$$

These boundary conditions are chosen for convenience: constant heat flux implies a layer of fixed thickness, and constant pressure resembles an open lower boundary (e.g. polar ocean underneath sea ice layer).

Non-dimensionalise the problem using scalings:

$$L \sim h, \quad t \sim \frac{h^2 \Omega}{\kappa}, \quad u \sim \frac{\kappa}{h \Omega}, \quad p \sim \beta^* \Delta C \rho_o g h$$

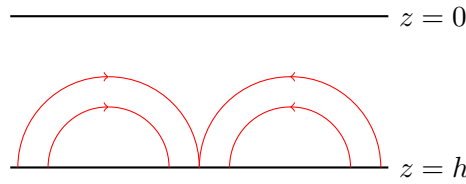
where  $\Omega = 1 + \mathcal{S}/\mathcal{C}$ . Here  $\beta^* = \beta - m\alpha$  which follows from using  $\Delta T = -m\Delta C$  to get  $\Delta\rho = \rho_0\beta^*\Delta C$ . The governing equations are then

$$\begin{aligned} \frac{\partial \theta}{\partial t} + \mathbf{u} \cdot \nabla \theta &= \nabla^2 \theta \\ \mathbf{u} &= -R_m [\nabla p + \theta \hat{\mathbf{z}}] \end{aligned}$$

where the Rayleigh number  $R_m$  in this case is

$$R_m = \frac{\rho_0 g \beta^* \Delta C \pi h}{\kappa \mu} \Omega$$

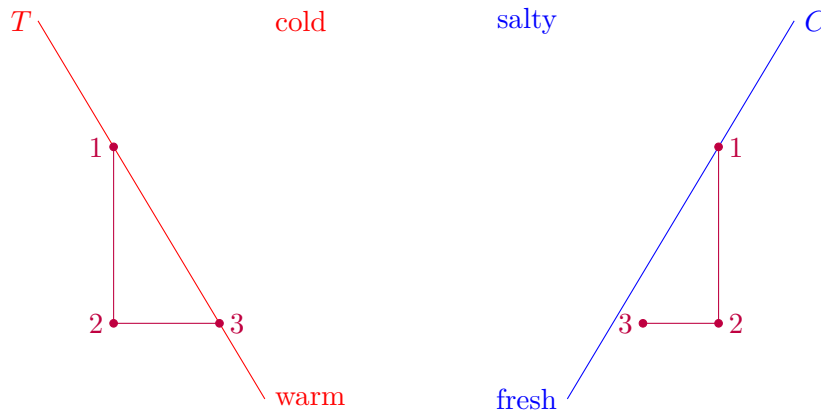
The solution resembles the convection rolls seen in section 3.2, except we now see only half of the rolls since the constant pressure boundary condition implies no tangential velocity component by Darcy's law.



The critical Rayleigh number is  $R_m = \pi^2$  at  $\alpha = \pi/2$ .

1. Convection is driven by compositional buoyancy, but limited by thermal dissipation.
2. Critical condition for convection is modified by phase change via  $\Omega = 1 + \mathcal{S}/\mathcal{C}$ . Large latent heat implies more unstable convection.

Consider temperature-concentration profiles:



A parcel of melt sea ice is cold and dense so descends ( $1 \rightarrow 2$ ), losing heat and salinity. The parcel rapidly thermally equilibrates ( $2 \rightarrow 3$ ) to the liquidus in the mushy layer, as thermal diffusion is fast. In contrast, diffusion of solute is slow so the concentration does not reach equilibrium. Salt must dissolve ice to reach equilibrium. Note that

$$\frac{\partial \phi}{\partial t} = -\frac{1}{\mathcal{C}} \frac{D\theta}{Dt}$$

so dissolution occurs when the temperature of a moving fluid parcel increases.

## 9 Poroelasticity

### 9.1 Flows of two-phase mixtures

Consider the flow/deformation of *both* the fluid and solid matrix, e.g. compaction of sediments, poroelastic deformation, melt generation in the mantle.

**Conservation of mass.** Treat both phases as separately incompressible and use the Boussinesq limit ( $\rho_s \approx \rho_l$ ). Conservation of mass is then expressed by

$$\begin{aligned} \phi_t + \nabla \cdot (\phi \mathbf{u}_s) &= \Gamma & \text{Solid} \\ (1 - \phi)_t + \nabla \cdot [(1 - \phi) \mathbf{u}_f] &= -\Gamma & \text{Liquid} \end{aligned}$$

where  $\phi$  is the solid fraction and  $\Gamma$  is the solidification rate. Note the average velocity

$$\bar{\mathbf{u}} = \phi \mathbf{u}_s + (1 - \phi) \mathbf{u}_f$$

satisfies  $\nabla \cdot \bar{\mathbf{u}}$  by above. Each phase itself is *not* separately divergence free (despite being separately incompressible).

**Conservation of momentum.** Assume Stokes flow for both phases, i.e.  $\text{Re} \ll 1$ . Conservation of momentum is expressed by

$$\begin{aligned} 0 &= -\nabla \cdot (\phi \sigma_s) + \phi \rho_s \mathbf{g} - \mathbf{h} & \text{Solid} \\ 0 &= -\nabla \cdot [(1 - \phi) \sigma_f] + (1 - \phi) \rho_f \mathbf{g} + \mathbf{h} & \text{Liquid} \end{aligned}$$

where  $\sigma_s, \sigma_f$  are the solid, fluid stress tensors and  $\mathbf{h}$  is the force exerted between phases.

Consider the limit where deformation of the solid phase is much more difficult than the fluid phase, e.g.  $\mu_f \ll \mu_s$ . Then

$$\sigma_f = -pI + \mu_s (\nabla \mathbf{u}_f + \nabla \mathbf{u}_f^T) \approx -pI$$

Take the interaction force to be

$$\mathbf{h} = c(\mathbf{u}_f - \mathbf{u}_s) + p \nabla \phi$$

Conservation of momentum in the fluid is then

$$0 = -\nabla \cdot [(1 - \phi)(-pI)] + (1 - \phi) \rho_f \mathbf{g} + c(\mathbf{u}_f - \mathbf{u}_s) + p \nabla \phi$$

Hence we have

$$c(\mathbf{u}_f - \mathbf{u}_s) = -(1 - \phi) [\nabla p + \rho_f \mathbf{g}]$$

We choose  $c = \mu_f(1 - \phi)^2/\Pi$  to recover Darcy's law. Then

$$(1 - \phi)(\mathbf{u}_f - \mathbf{u}_s) = -\frac{\Pi}{\mu_f} [\nabla p + \rho_f \mathbf{g}] \quad (18)$$

describes deformation of the solid matrix. Note that the fluid and solid matrix jointly support mechanical load. Consider the total stress

$$\begin{aligned} \sigma &= \phi \sigma_s + (1 - \phi) \sigma_f \\ &= \phi \sigma_s - (1 - \phi) pl \\ &= \phi(\sigma_s + pl) - pl \end{aligned}$$

Hence part of the isotropic solid stress is supported by the pore pressure. Introduce *Tarzaghi's effective stress*  $\sigma_e = \phi(\sigma_s + pl)$ . Then

$$\nabla \cdot \sigma = \nabla \cdot [\sigma_e - pl] = [\phi \rho_s + (1 - \phi) \rho_f] \mathbf{g}$$

**Poroelasticity.** We must choose a rheology to describe the solid. Consider moderate fluid stresses, a solid that deforms elastically (small deformation), e.g. aquifers, subglacial hydrology. In general, the effective stress can be written

$$\sigma_e = \sigma_e(\nabla \xi) = (K - \frac{2}{3}G) \nabla \cdot \xi \mathbf{I} + G(\nabla \xi + \nabla \xi^T)$$

where  $\xi$  is the deformation,  $K$  is the bulk modulus and  $G$  is the shear modulus. We can simplify this by using an isotropic, linear elastic model:

$$\sigma_e = E \frac{\phi_0 - \phi}{\phi_0}$$

Then

$$\begin{aligned} \nabla \cdot \sigma &= \nabla \cdot \sigma_e - \nabla p \\ &= -\frac{E}{\phi_0} \nabla \phi - \nabla p \\ &= \bar{\rho} \mathbf{g} \end{aligned} \quad (19)$$

**Applied load.** Consider a stress  $-\sigma_L$  applied to the solid matrix (e.g. porous plunger). Then

$$-\sigma_L = E \frac{\phi_0 - \phi}{\phi_0}$$

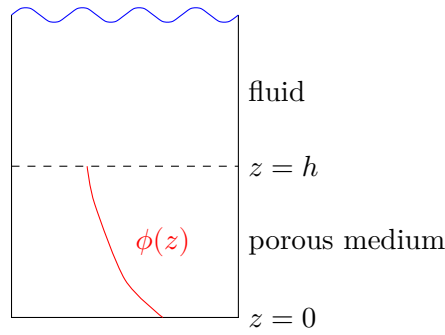
Rearranging for  $\phi$  we have

$$\phi = \phi_0 \left( 1 + \frac{\sigma_L}{E} \right)$$

## 9.2 Static compaction

Consider a model applicable to lake/sea floor sediments in which a static fluid flow compacts a porous medium due to pressure. No fluid flow implies pore pressure is hydrostatic

$$\frac{\partial p}{\partial z} = -\rho_f g$$



Using the poroelastic linear model (19) gives

$$\begin{aligned} -\frac{E}{\phi_0} \frac{\partial \phi}{\partial z} - \frac{\partial p}{\partial z} &= [\phi \rho_s + (1 - \phi) \rho_f] g \\ \Rightarrow -\frac{E}{\phi_0} \frac{\partial \phi}{\partial z} &= \phi (\rho_s - \rho_f) g \end{aligned}$$

Requiring  $\phi(h) = \phi_0$ , i.e. a stress free upper surface, gives

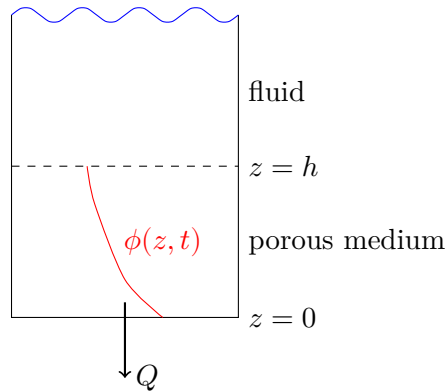
$$\phi = \phi_0 \exp \left[ \frac{\phi_0 \Delta \rho g}{E} (h - z) \right]$$

There is evidently a natural compaction length  $\delta$  given by

$$\delta = \frac{E}{\phi_0 \Delta \rho g}$$

### 9.3 Dynamic compaction

Now consider compaction with a fluid flux  $Q$  in the  $-\hat{z}$  direction. In this case  $\phi$  may be time-dependent.



Conservation of mass equations are

$$\begin{aligned} \phi_t + (\phi w_s)_z &= 0 \\ (1 - \phi)_t + [(1 - \phi) w_f]_z &= 0 \end{aligned} \tag{20}$$

Combining these equations and integrating gives the total mass flux, hence we have

$$\phi w_s + (1 - \phi) w_f = -Q \tag{21}$$



The poroelastic version of Darcy's law (18) gives

$$(1 - \phi)(w_f - w_s) = -\frac{\pi}{\mu_f} \left( \frac{\partial p}{\partial z} + \rho_f g \right) = -(Q + w_s)$$

where the last equality follows from (21). Substituting into conservation of mass in the solid (20) we find

$$\frac{\partial \phi}{\partial t} - Q \frac{\partial \phi}{\partial z} = \frac{\partial}{\partial z} \left[ \frac{\phi \pi}{\mu_f} \left( \frac{E}{\phi_0} \frac{\partial \phi}{\partial z} + \phi \Delta \rho g \right) \right]$$

Note  $\partial_t = Q = 0$  recovers the static result. Steady state compaction  $\phi_t = 0, w_s = 0$  gives

$$\begin{aligned} Q &= \frac{\pi}{\mu_f} \left[ \frac{E}{\phi_0} \frac{\partial \phi}{\partial z} + \phi \Delta \rho g \right] \\ \phi &= \frac{\mu_f Q \phi_0}{\pi E} \delta \left( 1 - e^{(h-z)/\delta} \right) + \phi_0 e^{(h-z)/\delta} \\ &\approx \phi_0 + \frac{\mu_f Q \phi_0}{\pi E} (h - z) \quad \text{for } h \ll \delta \end{aligned}$$

#### 9.4 Pore-pressure variations in a confined aquifer

Consider a confined aquifer with large lateral length and depth  $h(x, t) + H$  where the perturbation height  $h$  is posed as being due to pressure variation in the porous medium. We consider the influence of depth variations on a poroelastic medium constituting the aquifer. The problem is relevant to solid earth tides and how pressure is communicated subglacially. Take solid fraction

$$\phi = \Phi(x, t) \exp \left[ \frac{\delta \rho g \phi_0}{E} (h + H - z) \right]$$

where  $\Phi(x, t)$  is the solid fraction at the top of the aquifer. If

$$h + H \ll \frac{E}{\phi_0 \Delta \rho g}$$

then we may neglect vertical variations and linearise  $\phi(x, z, t) \approx \Phi(x, t)$ . Since over long lateral length scales the fluid velocity dominates  $u_f \gg u_s$ , we have

$$\int_0^{h+H} \phi \, dz = \text{const.} \implies (h + H)\Phi = H\Phi_0$$

from local solid mass conservation, where  $\Phi_0$  is the reference solid fraction. Now horizontal stress balance gives

$$\begin{aligned} \nabla \cdot \sigma_e \cdot \hat{x} - \frac{\partial p}{\partial x} &= 0 \\ \implies -\frac{E}{\phi_0} \frac{\partial \phi}{\partial x} &= \frac{\partial p}{\partial x} \end{aligned}$$

Consequently the modified Darcy's law becomes

$$(1 - \phi)u_f \approx \frac{\pi E}{\mu_f \phi_0} \frac{\partial \phi}{\partial x} = -\frac{\pi E}{\mu_f \phi_0} \frac{H}{(h + H)^2} \frac{\partial h}{\partial x}$$

Local fluid mass conservation is

$$\begin{aligned} \frac{\partial}{\partial t} [(h+H)(1-\phi)] + \frac{\partial}{\partial x} [(h+H)(1-\phi)u_f] &= 0 \\ \implies \frac{\partial h}{\partial t} &= \frac{\partial}{\partial x} \left[ \frac{\pi E}{\mu_f} \frac{H}{H+h} \frac{\partial h}{\partial x} \right] \end{aligned}$$

but for typical values deformation is small,  $h \ll H$ , hence

$$\frac{\partial h}{\partial t} \approx \frac{\pi E}{\mu_f} \frac{\partial^2 h}{\partial x^2} \quad (22)$$

Stress balance at the top surface is

$$\begin{aligned} -p + \hat{\mathbf{z}} \cdot \boldsymbol{\sigma}_e \cdot \hat{\mathbf{z}} &= -p_0 \\ -p + E \frac{\phi_0 - \phi}{\phi_0} &= -p_0 \\ p - p_0 &= E \frac{\phi_0 - \phi}{\phi_0} = E \frac{h}{H} \end{aligned} \quad (23)$$

where  $p_0$  is the *overburden pressure*. Combining (22) and (23) gives

$$\frac{\partial p}{\partial t} = \frac{\pi E}{\mu_f} \frac{\partial^2 p}{\partial x^2}$$

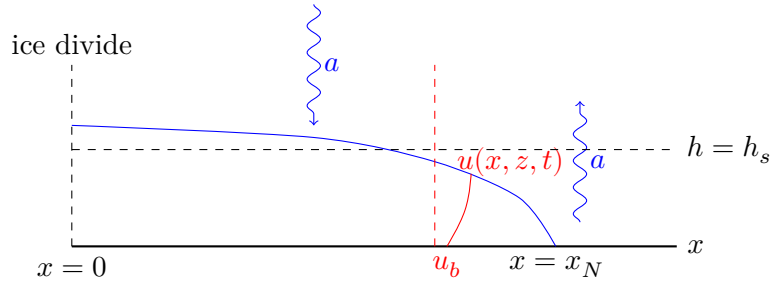
## 10 Ice sheets

### 10.1 Terrestrial ice sheets

In this section we consider ice sheets lying over land (e.g. Greenland). Ice sheets are fed by snowfall at high latitudes and are terminated on land by surface melting/ablation. They flow as a viscous current on timescales longer than months.

#### 10.1.1 Two dimensional model

Consider a two dimensional model of an ice sheet with a uniform (flat) bedrock. We consider a region between the *ice divide*  $x = 0$  where ice separates and the *nose*  $x = x_N$  where the ice sheet terminates.



We suppose there is an accumulation/ablation rate  $a = a(x, t, h)$  (consider only  $a_N < 0$ ) which is positive above the *snow line*  $h = h_s$  and negative below it. The governing equation for the ice sheet depth  $h$  is

$$\frac{\partial h}{\partial t} + \frac{\partial q}{\partial x} = a$$

where  $q$  is the mass flux and we assume it is given by

$$q = -\frac{\rho g}{3\mu} h^3 \frac{\partial h}{\partial x}$$

as for a gravity current. The boundary conditions are

$$\begin{aligned} q &= 0 & \text{at } x &= 0 \\ q &= 0 & \text{at } x &= x_N \\ h &= 0 & \text{at } x &= x_N \end{aligned}$$

Suppose the accumulation/ablate rate  $a \sim A$  has characteristic amplitude. Let  $h \sim h_s, t \sim h_s/A, x \sim \left(\frac{\rho g h_s^4}{3\mu A}\right)^{1/2}$ . The dimensionless governing equation is then

$$\frac{\partial h}{\partial t} = \frac{\partial}{\partial x} \left( h^3 \frac{\partial h}{\partial x} \right) + a \quad (24)$$

### 10.1.2 Dynamics near the front

At the front of the ice sheet  $h(x_N(t), t) = 0$ . Differentiating with respect to  $t$  gives

$$\frac{\partial h}{\partial x} \dot{x}_N + \frac{\partial h}{\partial t} = 0$$

at the nose  $x = x_N(t)$ . Hence *near* the nose

$$h_t \approx -\dot{x}_N h_x$$

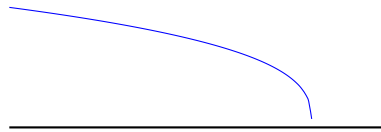
The dynamics and shape of the front are determined by which terms of the governing equation (24) form the dominant balance. Consider three possibilities for the ablation rate at the nose  $a_N$ :

1. No ablation,  $a_N = 0$ : only one possible balance

$$\dot{x}_N h_x \sim (h^3 h_x)_x = \frac{1}{4} (h^4)_{xx}$$

The shape of the nose is then

$$h \sim (x_N - x)^{1/3}$$



2. Ablation at the nose  $a_N < 0$ :

a) advancing nose  $h_t > 0$ : try  $h^4/x^2 \sim -a_N$ . Then

$$\begin{aligned} h &\sim (-a_N)^{1/4} (x_N - x)^{1/2} \\ h_t &\sim -\dot{x}_N h_x \sim \frac{\dot{x}_N (-a_N)^{1/4}}{(x_N - x)^{1/2}} \\ &\rightarrow \infty \text{ as } x \rightarrow x_N \end{aligned}$$

hence the system is unbalanced. We must therefore use the balance as in 1. above, so  $h \sim (x_N - x)^{1/3}$  again.

b) steady nose  $h_t = 0$ : only balance is  $q_x \sim a_N$  as above which gives

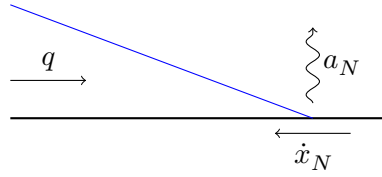
$$h \sim (-a_N)^{1/4}(x_N - x)^{1/2}$$

which is now balanced as  $h_t$  vanishes, i.e. flow towards the front is balanced by ablation.

c) retreating nose  $h_t < 0$ : system is again unbalanced if we try  $q_x \sim (h^3 h_x)_x \sim a_N$  but we can instead balance  $h_t \sim a_N$ . Then

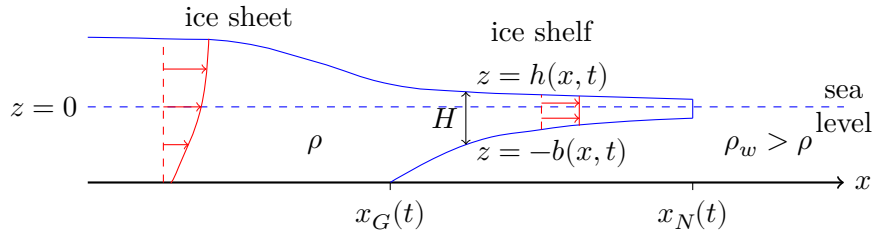
$$\begin{aligned} h_t &\sim -\dot{x}_N h_x \sim a_N \\ \Rightarrow h &\sim \frac{-a_N}{-\dot{x}_N}(x_N - x) \end{aligned}$$

i.e. a linear profile at the nose.



## 10.2 Marine ice sheets

Marine ice sheets sit on bedrock below sea level and detach (i.e. float) at the *grounding line* to form an ice shelf.



We model the ice shelf as extensional flow with an almost uniform profile and total ice depth  $H = h + b$  with height  $h(x, t)$  above sea level and depth  $-b(x, t)$  below sea level. The ice shelf has density  $\rho$  and the surrounding water has density  $\rho_w > \rho$ , hence the shelf floats.

### 10.2.1 Ice shelf height

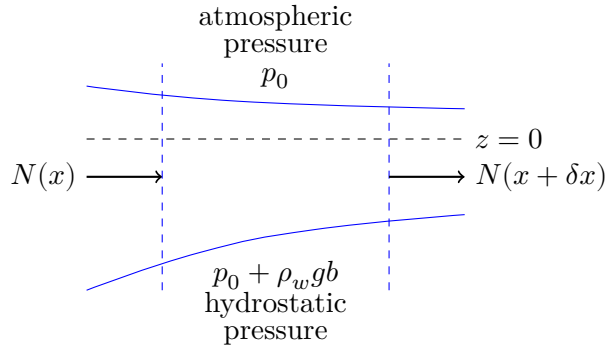
Ignore bending moments and apply Archimedean buoyancy to find the height above sea level: weight of the ice shelf must balance the buoyant force provided by the surrounding water.

$$\begin{aligned} \rho g H &= \rho_w g (H - h) \\ \Rightarrow h &= \frac{g'}{g} H \end{aligned} \tag{25}$$

where  $g' = \frac{\rho_w - \rho}{\rho_w} g$  is the *reduced gravity*. Note that typically  $h \sim H/7$ . The result (25) may be referred to as the *flotation condition*.

### 10.2.2 Governing equations

**Stress balances.** Consider the stress balances on a small section of ice shelf.



The vertical balance is

$$\frac{\partial}{\partial z} \sigma_{zz} = \rho g \implies \sigma_{zz} = -p_0 - \rho g(h - z)$$

But for a Newtonian fluid  $\sigma_{zz} = -p + 2\mu \frac{\partial w}{\partial z}$ . Hence, choosing  $p_0 = 0$  wlog,

$$p = \rho g(h - z) + 2\mu \frac{\partial w}{\partial z}$$

Then

$$\begin{aligned} \sigma_{xx} &= -p + 2\mu \frac{\partial u}{\partial x} \\ &= -\rho g(h - z) - 2\mu \frac{\partial w}{\partial z} + 2\mu \frac{\partial u}{\partial x} \\ &= -\rho g(h - z) + 4\mu \frac{\partial u}{\partial x} \end{aligned}$$

using  $u_x + w_z = 0$  by incompressibility. The forces  $N(x)$  are then

$$\begin{aligned} N &= \int_{h-H}^h \sigma_{xx} dz \\ &= -\frac{1}{2} \rho g H^2 + 4\mu H \frac{\partial u}{\partial x} \end{aligned} \tag{26}$$

Horizontal balance on a section of shelf then balances the stress  $N_x$  with the stress from the horizontal component of the buoyancy force:

$$\begin{aligned} \frac{\partial N}{\partial x} + \rho_w g b \frac{\partial b}{\partial x} &= 0 \\ 4\mu \frac{\partial}{\partial x} \left( H \frac{\partial u}{\partial x} \right) &= \rho g' H \frac{\partial H}{\partial x} \end{aligned} \tag{27}$$

This is an elliptic equation for  $u$  given  $H$ .

**Conservation of mass.** Conservation of mass is expressed by

$$\frac{\partial H}{\partial t} + \frac{\partial}{\partial x} (Hu) = 0 \tag{28}$$

This is a hyperbolic equation for  $H$  given  $u$ .

**Boundary conditions.** At the nose  $x = x_N$  we have two boundary conditions

$$\begin{aligned} N &= -\frac{1}{2}\rho_w g b^2 \\ \dot{x}_N &= u_N \end{aligned} \tag{29}$$

where the first specifies the force on the nose as being provided solely by the horizontal buoyancy force, and the second defines the nose velocity  $u_N$ .

At the grounding line  $x = x_G$  we have two boundary conditions

$$\begin{aligned} H &= H_G \\ Hu &= q_G \end{aligned}$$

where the first specifies that the floating depth of the ice equals the depth of the sea at the grounding line, and the second enforces a constant flux into the ice shelf.

**Example.** Let  $H = H_0, q = q_0 = Hu$  at  $x = 0$ . Integrate (27) and use the flotation condition (25) combined with (29) and (26) at the nose to eliminate the integration constant, hence

$$\frac{\partial u}{\partial x} = \frac{\rho g'}{8\mu} H$$

where  $g' = \Delta\rho g/\rho$ . The solution is then

$$\begin{aligned} u &= \frac{q_0}{H_0} + \int_0^x \frac{\rho g'}{8\mu} H(\xi, t) d\xi \\ &= \frac{q_0}{H_0} + \frac{\rho g' q_0}{8\mu} (t - t_0) \end{aligned} \tag{30}$$

where  $t_0(x)$  is the time at which fluid now at  $x$  left  $x = 0$ . Note: this is effectively a switch between Eulerian and Lagrangian perspectives of mass conservation. Therefore integrating (30) gives

$$x = \frac{q_0}{H_0} (t - t_0) + \frac{\rho g' q_0}{16\mu} (t - t_0)^2 \tag{31}$$

Now from mass conservation (28) we have

$$\begin{aligned} H_t + uH_x &= -Hu_x \\ \implies \frac{DH}{Dt} &= -\frac{\rho g'}{8\mu} H^2 \\ \implies \frac{1}{H} - \frac{1}{H_0} &= \frac{\rho g'}{8\mu} (t - t_0) \end{aligned} \tag{32}$$

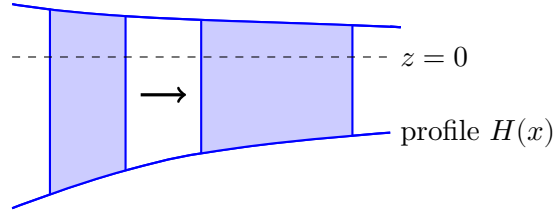
Combine (31) and (32) to give the profile of shelf thickness

$$H = H_0 \left[ 1 + \frac{\rho g'}{4\mu} \frac{H_0^2 x}{q_0} \right]^{-1/2}$$

and from (31)

$$x_N(t) = \frac{q_0}{H_0} t + \frac{\rho g'}{16\mu} t^2$$

The first term is due to extrusion of the ice sheet into the ice shelf, whilst the second term is due to buoyancy-driven extension of the ice sheet.



### 10.3 Grounding line

Consider the dynamics of grounding lines. The flotation condition (25) is first satisfied by the shelf at the grounding line:

$$h(x_G(t)) = \frac{\rho_w - \rho}{\rho} b(x_G(t))$$

This condition must be satisfied for all time, hence we can differentiate

$$\dot{x}_G \left. \frac{\partial h}{\partial x} \right|_{x_G^-} + \left. \frac{\partial h}{\partial t} \right|_{x_G^-} = \frac{\rho_w - \rho}{\rho} \dot{x}_G b'(x_G)$$

Rearranging gives

$$\left[ \frac{\Delta \rho}{\rho} b'(x_G) - \left. \frac{\partial h}{\partial x} \right|_{x_G^-} \right] \dot{x}_G = \left. \frac{\partial h}{\partial t} \right|_{x_G^-} = - \left. \frac{\partial q}{\partial x} \right|_{x_G^-} \quad (33)$$

We require an additional condition, that normal forces are balanced at the grounding line:

$$[N]_{x_G^-}^{x_G^+} = 0$$

In the shelf,

$$\begin{aligned} N &= -\frac{1}{2} \rho g H^2 + 4\mu H \frac{\partial u}{\partial x} \\ &= -\frac{1}{2} \rho g H^2 + \frac{1}{2} \rho g' H^2 \\ &= -\frac{1}{2} \rho_w g b^2 \end{aligned}$$

This is the hydrostatic force of the ocean pushing on the sheet at the grounding line. The presence of the shelf has no influence on the 2D ice sheet. In the sheet,

$$\begin{aligned} N &= \int_{-b}^h \sigma_{xx} dz \\ &= -\frac{1}{2} \rho g H^2 + 4\mu \int_{-b}^H \frac{\partial u}{\partial x} dz \\ &= -\frac{1}{2} \rho g H^2 + 4\mu \left[ \frac{\partial q}{\partial x} - u(h) \frac{\partial h}{\partial x} - u(-b) \frac{\partial b}{\partial x} \right] \end{aligned}$$

with parabolic velocity profile (following from lubrication theory with no slip at bottom, no stress at surface)

$$u = -\frac{\rho g}{2\mu} \frac{\partial h}{\partial x} (z+b)(2H - (z+b))$$

Then

$$N = -\frac{1}{2} \rho_g H^2 + 4\mu \left[ \frac{\partial q}{\partial x} + \frac{\rho g}{2\mu} H^2 \left( \frac{\partial h}{\partial x} \right)^2 \right]$$

From (33) we then find

$$\dot{x}_G = \frac{\frac{\rho g}{2\mu} H^2 h_x^2 - \frac{\rho g'}{8\mu} H^2}{\frac{\Delta\rho}{\rho} b'(x_G) - h_x|_{x_G^-}}$$

In steady state  $\dot{x}_G = 0$  we have

$$\begin{aligned} \frac{\rho g}{2\mu} h_x^2 &= \frac{\rho g'}{8\mu} \\ \Rightarrow \frac{\partial h}{\partial x} &= -\frac{1}{2} \sqrt{\frac{\Delta\rho}{\rho}} \end{aligned}$$

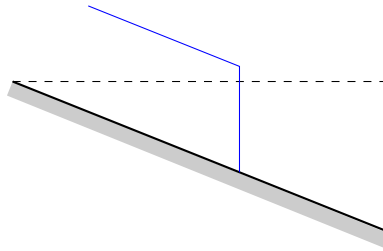
The flux is then

$$q = -\frac{\rho g}{3\mu} H^3 h_x = \frac{\rho g}{6\mu} \sqrt{\frac{\Delta\rho}{\rho}} H^3 = q(H)$$

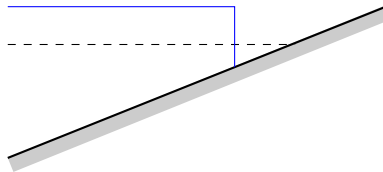
### 10.3.1 Grounding line stability

See Weertman 1974 for further details on this topic. The stability of the grounding line is determined by the slope of the ground:

- **Prograde slope:** if  $x_G$  advances,  $H$  increases, hence  $q \sim H^3$  increases. Thus the *sheet* thins, so  $x_G$  retreats. This situation is stable.



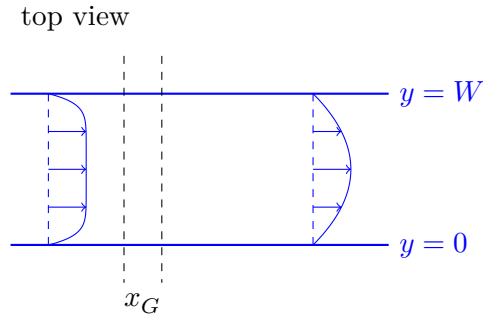
- **Retrograde slope:** if  $x_G$  advances,  $H$  decreases, hence  $q \sim H^3$  decreases, so the *sheet* thickens, thus  $x_G$  advances further. This situation is unstable.



### 10.3.2 Buttressing

Thus far we have only considered the vertical and extensional directions of the ice shelf. Buttressing is resistance to flow at the side walls of the ice shelf. In the ice *sheet*, vertical shear stresses dominate. In the ice *shelf*, horizontal shear stresses dominate.





In the shelf,

$$u = -\frac{\rho g}{2\mu} \frac{\partial h}{\partial x} y(W - y)$$

The additional horizontal force in the  $x$ -direction from buttressing is given by

$$\begin{aligned} -B &= \frac{1}{W} \int_{x_G^+}^{x_N} H \mu \left. \frac{\partial u}{\partial y} \right|_{y=0}^{y=W} dx \\ &= -\frac{1}{2} \rho g H^2 \Big|_{x_G^+} \end{aligned}$$

Finally we get

$$\dot{x}_G = \frac{\frac{\rho g}{2\mu} H^2 h_x^2 - \frac{\rho g'}{8\mu} (H^2 - H_+^2)}{\frac{\rho_w}{\rho} b'(x_G) - \left. \frac{\partial h}{\partial x} \right|_{x_G^-}}$$

where  $H_+ = \lim_{x \rightarrow x_G^+} H$  in the shelf dominated by horizontal shear stresses. Note that the buttressing force tends to *advance* the grounding line position, so can mitigate instability/retreat of the grounding line.