Olga Sergienko on understanding dynamics of ice streams and ice shelves:
You need to do the conservation of energy (well understood, not discussed here), mass and momentum. The basic ingredients are:

1) Ice is a non-Newtonian fluid (acts like honey) and is incompressible
2) Newton's second law is written with "deviatoric" stresses
3) Nonlinearity comes in from Glen's flow law (strain rate = A(T) * effective stress ^(n-1)) where <u>viscosity</u> itself is a function of **strain rate** and **temperature**
4) Scaling: Ice is/has…
   i) much wider (W) than thick (H)
   ii) generally slow velocities
   iii) high pressure

   Which implies that kinetic energy negligible to potential energy (low Froude Num.) and inertia may be neglected

Steps to the Shallow Shelf Approximation:

- Starting from Navier-Stokes without inertial terms
- Boundary conditions come from cartoon: surface (no force, kinematic), bed (pressure, frozen, sliding, floating i.e. hydrostatic)
- We obtain the full stokes equations + incompressibility: 4 unknowns & 4 eqns. -> solvable but far too messy
- simplification by "asymptotic analysis": small H/L + non-dimensional perturbation theory, which yields zero and higher order approximations (still a bit messy…)
- finally ending up with elliptic SSA (zero order without vertical shear stress and with vertical homogenous velocities)

In other words: "On an ice shelf vertical strain rates become zero and archimedes principle can be used to calculate buoyancy"

Same dynamics can be applied to ice streams with low basal friction/ basal shear stress. This is the basis of the shelfy stream model

Summary: Ice Streams, Lakes, Calving and Fracture

- Ice streams observed to respond to sub-daily tidal modulations and are capable of responding to high frequency forcing
- High frequency stick-slip behaviour is observed on Whillans, which is slowing down, with large variation near "sticky spot." Static vs. Dynamic friction
- Subglacial lakes are observed under the ice sheet and ice streams, capable of storing and releasing large quantities of water.
- Calving incorporates a range of types and processes from rapidly turning over tidewater margins to big ice shelf rifts
- Important concepts are the calving rate, calving flux and calving "laws" (which include Brown's water depth, Alley's simple law and Van der Veen/Vieli height above buoyancy criteria)
- Fracture is characterised in 3 modes, crevasse models such as Nye and Weertman use only mode 1 fracture to determine depth.
- Water added to a fracture enhances propagation
- Statistical approach of Bassis parameterises calving rate as a probability of iceberg initiation proportionally to the calving rate, combining water depth and height above buoyancy "laws".

Adrian Jenkins on Ice shelf ocean interactions:
As we know, ice shelves are the floating margins of an ice sheet. Mass discharge of icebergs and basal melt water are important contributors of freshwater to the ocean, especially in the southern

hemisphere and at depth. The thickness of the ice shelf is dependent on accumulation, vertical strain and most importantly melt/freeze.

- When ice and seawater come into contact phase changes occur if the water is not at the freezing point.
- The rate at which the phase change occurs is determined by diffusion across the turbulent boundary layer. The interface between ice and seawater adjusts rapidly to equilibrium.

- Freezing point is linear function of salinity and pressure
- The interfacial sub-layer, where turbulence is suppressed has most change occurring across it (~75% heat, ~99% salt), by molecular diffusion. It's very thin.

Understanding Basal Melting:
It is important to understand how melting ice affects the heat and salt balance within the seawater, especially under a shelf. Adrian described an experiment for obtaining heat flux using a hot water drill through the ice shelf to collect Temperature (T) and Salinity (S), and radar to measure sub-cm changes in ice thickness). They assumed that vertical velocity doesn't vary – displacement at bottom is strain due to melting.

- Transfer coefficients for T, S used in diffusivities to compute heat flux were optimized using time series (constant ratio between T,S was assumed). This allowed determination of a combined transfer coefficient (for T and S) that could be used more generally.

Water motion due to phase changes:
Phase changes due to T, P, or S cause downwelling or upwelling. When seawater interacts with ice, seawater cools and dilutes, and properties evolve along straight cooling/dilution lines in the standard phase diagram shown in Adrian's lecture.

At some defined pressure, you compare the slope for cooling/dilution lines and isopycnal contours. This generates a line for determining downwelling (above the line) or upwelling (below the line) due to melting. In polar regions, melting leads to upwelling.

This process is important for ice shelves where the grounding line is at depth. The freezing temperature for seawater decreases for each 1000m drop in depth (increase in P). Then, as ice melts at depth and meltwater is buoyant, it rises to shallower depth and freezes. This "ice pump" is a mechanism of energy transport within the system, by lowering the potential energy without requiring an external input of energy. Frazil ice forms in suspension when supercooled seawater freezes to the underside of the ice shelf.

Buoyancy driven flows: "plumes"
Plumes are characterized as/ by:

- Buoyant & turbulent flow that entrains fluid from surroundings
- Entrainment rate & heat flux into ice depends on plume velocity (U)
- Plume velocity depends on density difference between plume and surroundings

Conservation of mass, -momentum, -heat, -salt give four eqns. which can be solved by using the linearized eqn. of state an entrainment law and the earlier discussed ice-ocean thermodynamics.
The plume will evolve as follows:
Melting increases buoyancy -> plume flows upslope along the ice base and grows –> entrainment is less effective as plume grows -> melting becomes weaker and may stop -> the plume might leave the ice base when it reaches a level of neutral density. Moreover we discussed many other funny effects e.g. due to change of the pressure melting point, slope of the ice base dependency, subglacial runoff, rotational effects.

Finally, one of the take home messages was: Melt rate ($\dot{m}$) depends linearly on temperature and velocity in mixed layer, which implies for different cases:

i) classical (2-D) plume: $U \propto \sqrt{T_{far}} \rightarrow \dot{m} \propto T_{far}{}^{1.5}$

ii) Geostrophic 3-D circulation: $U \propto T_{far} \rightarrow \dot{m} \propto T_{far}{}^{2}$

iii) Buoyancy (velocity) externally controlled (e.g. subglacial melt, tides) yields linear melting with far field temperature (T_{far})