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R. Hindmarsh: Ice Sheets

What is an ice sheet?

There are two different types of ice sheets: (1) Greenland type (snowfall/accumulation at higher elevations, ablation at the edges), often has fjord-terminating outlet glaciers (2) Antarctica type (most of the ice is in an accumulation zone, ice is lost by calving/submarine melting at the sea-terminating boundary). Ice sheets that terminate in the ocean and have a bed that is below sea level (i.e., West Antarctica) are most susceptible to rapid change.

Complexities

Ice sheets flow slowly (meters per year) at high elevations. As the ice funnels into narrow outlets, the flow can become rapid and streaming (a few km per year). Complex flow features include ice streams and ice shelves.

Ice sheet configurations: past and future

Although ice sheets are usually considered to evolve slowly over millennia, recent and geologic evidence indicates that changes can occur over centennial time scales. These rapid changes result from instabilities in glaciers that terminate in the ocean. Some studies have suggested that variations in surface melting and associated basal motion can influence ice sheet stability, but this is most likely a very minor effect. A recent model by Pollard and DeConto suggests that the Antarctic Ice Sheet can maintain at least 3 different steady states: extreme interglacial (no ice shelves), modern interglacial (some ice shelves), and full glacial (ice extends over entire continental shelf).

Why do we care about forecasting ice sheets?

Global sea level rise is 2-3 mm/yr at present. Greenland contains 7 m sea level equivalent, and is more directly sensitive to climate forcing. West Antarctica contains 5 m sea level equivalent and is more indirectly controlled. The Antarctic Peninsula is the most sensitive area of Antarctica, with 0.5 m sea level equivalent.

What do we need to know to forecast, and what are the outstanding obstacles?

Need to know accumulation, ice velocity, ice thickness, and the rate of thickness change. Difficulties associated with modeling ice sheet evolution include the observation that retreat rates can be very fast relative to typical model time steps, there is lack of data to validate models, and it is difficult to model grounding line retreat, iceberg calving, and subglacial processes. At present no ice-sheet model predicts both retreat of the Antarctic ice sheet since the LGM and represents current variability.

What do ice-sheet models need?

In order to model ice-sheet thickness, we need the boundary value of mass-balance forcing, as well as heat flux, and we need to describe the ice flow.

What do ice-sheet models try to do?

1. Describe the evolution of ice sheets over tens of thousands to millions of years. For example, the study by Pollard and DeConto (2009) models the evolution of Antarctica over the past 5 Myr and they identify at least three ice-sheet modes: full glacial, modern interglacial, and extreme interglacial. However, models of ice-sheet evolution may need improvements in order to match observations. For example, Vinther and others (2009) find that models of Greenland are inconsistent with ice-sheet history inferred from oxygen isotopes.

2. Help describe paleoclimate phenomena. For example, MacAyeal presented a binge-purge mechanism of internal dynamics for Heinrich events. The ice sheet grows, and once it is thick enough the ice can begin sliding and loses mass until the surge can no longer be sustained and the cycle is reset. This model can match the ~7,000-10,000 year periodicity of the Heinrich events. We also discussed the model of Payne (1995), which discussed limit cycles in the basal thermal regime.

3. Couple to global circulation models. For example, Otto-Bliesner and others (2006) studied Arctic climate during the last interglacial.

4. Predict observations. Data can be used directly in models, or a model can be used to match something we observe in order to infer something about the forcing or the governing physics. Inferences from ice-sheet data can be made by solving a minimization problem, or by solving an inverse problem. For example, Nick and others (2009) simulated Helheim acceleration, thinning, and retreat initiated by a perturbation at the terminus.

5. Calculate balance velocities and characterize the large-scale modern state. We don't know the viscosity parameter well, but we can use mass conservation to diagnose the present state of the ice sheet and for the initialization of ice-sheet models. For example, Bamber and others have done this for Antarctica.

6. Sensitivity analysis. For example, Payne and others (2004) have modeled the downstream effects on ice streams – how changes in grounding-line position change the glacier geometry. We discussed the novel formulation of Schoof (2007) to describe the flux across the sheet-shelf transition.

7. Forecasting Greenland and Antarctica evolution. Unfortunately, a model that fits paleoclimate data and modern data does not exist. We know that rapid changes can occur, and these processes need to be properly included in ice-sheet models. We also need to consider the link between ice sheets and the rest of the climate system; developing a sufficiently coupled model of the ice-atmosphere-ocean remains an open challenge.

F. Stranneo: Warm Waters at High Latitudes

Polar Heat Transport
There is a radiation gain in the tropics and loss at high latitudes. As a result, there is poleward heat transport by the ocean-atmosphere system. In the Northern Hemisphere, the Atlantic does the bulk of the heat transport. Ocean-driven poleward heat transport has no one driver. It's associated with the warm to cold conversion of water masses poleward, and is achieved by the combination of horizontal and vertical gyres (driven by wind, heat, freshwater fluxes and tides).

Hemispheric differences
A comparison of 50°N and 50°S shows that the Northern Hemisphere ocean is much warmer than in the south. This is due largely to the continents in the NHem, which allow heat to climb farther north because the flow is confined by the narrow topography.

Pacific/Atlantic differences
In the South Atlantic, heat transport is north (equatorward), whereas the South Pacific transports heat poleward. The Atlantic has two clockwise subtropical gyres and one anticlockwise subpolar gyre.

North Atlantic
Heat is moved from boundary currents to the interior by eddies (and then lost to the atmosphere). Densest waters are formed in the interior. Cooling is due to direct surface heat loss to the atmosphere and shedding eddies into the interior.

Nordic Seas
The same idea holds in the Nordic Seas, except the topography is more complicated. A sill impacts the circulation/overflow. The sill partially "blocks" the warm inflow to the Nordic Seas.

Arctic Ocean
There are two main deep basins, broad shelves, and except for Fram Strait they are connected by shallow straits. Arctic is mostly seasonally ice covered. The cold Arctic halocline keeps the warm NADW separated from the surface where it could melt sea ice. Waters are cooled and freshened in the Arctic. Seasonal variability: WGC (Arctic outflow) is faster and fresher in spring.

Variability: Interannual to multidecadal
Variability in convection can lead to variability in overturning circulation and poleward heat transport. Global climate is sensitive to changes in dense water formation (can be seen with changes in freshwater discharge which, in extreme, can shutdown MOC).

North Atlantic Oscillation (NAO)
The NAO is a large-scale alteration of atmospheric mass, accounting for >1/3 SLP variance. It is the dominant mode of atmospheric variability in the North Atlantic. The positive phase brings more storms across the Atlantic (storm track shifts closer to Greenland). NAO has weekly variability, but is considered decadal.