Summary of lectures by Svein Oesterhus – Saturday, 12 June 2010

One Hundred Years in the Nordic Seas

Norway has maintained an active observational physical oceanography program throughout the 20th century, largely in support of the fisheries that are critical to their economy and culture. The 1909 book *The Norwegian Sea* by Bjorn Helland-Hansen and Fridtjof Nansen gives the initial comprehensive characterization of the hydrography (temperature, salinity, and primary currents) of the Nordic seas. Since then physical oceanographers have:

- made major improvements in instrumentation, going from inverting thermometers to modern CTDs and current meters;
- measured fluxes of heat, mass, and salt; and
- improved our knowledge of the variability of the Nordic seas.

There are many long time series throughout the Nordic seas, and they show coordinated variability on decadal timescales. Broadly speaking, we observe a slow warming of the Nordic Seas from 1900-1960, followed by a cooling period from 1960-1980, and warming since 1980. Salinity variations are superimposed on this temperature variation. The Great Salinity Anomaly was a surface fresh water pulse in the late 1960s and early 1970s. Since then, we have observed a steady freshening in an ever-thicker surface layer. Finally, there are century-scale variations in sea ice extent: in the 1750s, the southern-most extent of the Arctic sea ice was north of Svalbard; by the end of the 19th century, the sea ice reached almost as far south as Tromso; today, the sea ice has once again retreated almost as far north as was seen in the earliest records.

The steady warming of the deep water is one of the most important changes observed in the Nordic seas. The mechanism for this warming is critically dependent on the geographical context of the Nordic Seas. The Greenland and Norwegian seas are 4000m deep, but isolated from both the Arctic and Atlantic basins by very high ridges. This means that cold Arctic water spills over the ridge, but all other forcing of the water in the Greenland and Norwegian Seas must come from directly above. Before 1970, it appeared that wintertime convection mixed the Arctic overflow water throughout the full depth of the Greenland Sea to create very cold Greenland Sea Deep Water. Since 1970, the wintertime convection cells have been shallower, rarely penetrating below 1500m. The intermediate and deep water are therefore no longer being mixed with the very cold surface forced water, and the deep water has been effectively isolated from all external forcing. In this very quiet context, geothermal heating, a forcing mechanism that is usually considered negligible by oceanographers, may be very important. The geothermal heat flux in this area is approximately 60 mW/m^2; to understand what a weak forcing this is, recall that the average surface insolation is over 200 W/m^2, more than 1000 times larger. Geothermal heating is efficient at driving convection, as the heat is supplied at the bottom and it can create classical Rayleigh-Bernard overturning cells. The evidence for geothermally driven convection is constant salinity and potential temperature in the deep Greenland Sea. In addition, the potential temperature in the deep water is increasing at a rate that is consistent with an analytical model of geothermal convection. Geothermal convection may be important for longer climate timescales or during ice ages. These overturning cells reach higher and higher into the water column, and will require approximately 1000 years to occupy the full depth of the Greenland Sea. Once the geothermally-heated water reaches the surface, it could break up the overlying sea ice, allowing heat loss to the atmosphere and open ocean convection filling the full water depth. Once the geothermal heat is vented to the atmosphere, the sea ice will reform, allowing the process to restart its approximately 1000yr oscillation.

Thermohaline Circulation

Theormohaline circulation and meridional overturning circulation are terms that are used to mean a wide variety of things in oceanography. Wunsch (2002) identified at least seven different definitions. In this lecture, we adopted the definition: “circulation driven by density and pressure differences in the deep ocean”. In the Nordic Seas, it is observed as transports through a few key water passages: the Fram Strait, the Barents Sea, and over the Greenland-Scotland Ridge. These constrictions allow budgets to be observed and constrained using moorings and repeated hydrographic sections that cover almost all the available routes for water. No trend has been observed in the volume flux in the thermohaline circulation in the North Atlantic.

The other key location for deep water formation in the thermohaline circulation is the Southern Ocean. The Weddell Sea and the Greenland Sea produce the densest water in deep ocean, and act to drive the oceanic “conveyor” at the two poles. As in the Nordic Seas, there is a long history of Norwegian observing in the Southern Ocean, beginning from the Norvegia Expedition in 1927.

Deep water is formed by two mechanisms in the Southern Hemisphere. The first is open ocean convection, the same mechanism used in the Nordic Seas. The most famous example of this was in the Weddell Polynya in the 1970s. However, the Weddell Sea is usually covered by sea ice in the wintertime, preventing the necessary ocean-to-atmosphere flux for this. The second mechanism is shelf convection, in which the brine rejected from freezing sea ice combines with heat loss to the atmosphere to form High Salinity Shelf Water. Further cooling makes this water even denser, and it overflows into the deep ocean as Antarctic Bottom Water. The source and product waters can be seen clearly along a mixing line in Temperature-Salinity space, and Norway is developing systems for observing and monitoring this deep water formation process using ice-moored instruments.

Summary of Olga Sergienko's Lectures -- Saturday, June 12, 2010

Ice Shelf Collapse:

Over the past decade, we have observed two large ice shelf collapses on the Antarctic continent. On March 20, 2002, the Larsen B ice shelf collapsed over a very short time period. Modelling work done by Scambos et al., 2000, proposes that when a crevasse reaches a certain size (14-22 m deep, ~30m wide) and is approximately 90% filled with water, its threshold is breached, and it will catastrophically crack, setting off a calving event. The cause of Larsen B event, then, has been attributed to the large amount of meltwater observed over the shelf previous to its collapse.

The collapse of the Wilkins Ice Shelf in February/March of 2008 is less well understood. Satellite imagery shows no evidence of surface melt or widespread crevasses before the breakup. However, over a short period of time, more than 50 % of the original ice shelf was quickly converted into a blue mélange, rapidly expanding seaward faster than 30 cm/s, creating an impressive domino effect.

MacAyeal et al., 2008, suggested the following mechanism for the chain reaction: As If an initial perturbation is felt (say by a micro-tsunami hitting the shelf), as the ice goes up and breaks off, it is replaced by water, which is a denser material. Gravitational potential energy is therefore liberated, and the next piece of ice is broken off. Because the Wilkins shelf was made up of relatively fragile, compacted snowfall, a small forcing may have amplified to large effect.

For another possible reason for the breakup, we look to the Patagonian coast. The breakup of the Wilkins shelf was concurrent with anomalously high storm activity off of Patagonia. Perhaps long-wave infragravity waves made their way south and into the cavity beneath the Wilkins shelf, causing the shelf to resonate in a manner analogous to the Tacoma Narrows Bridge.

Is there evidence that ice shelves feel such waves? Indeed, an array of seismometers distributed along the Ross shelf shows that most of the recorded activity was detected in the long wave. The narrow opening between the shelf and sill probably deflects shorter wavelengths.

Inverse Modeling:

When dealing with a modeling problem, one is privy to two pieces of information: observations and a model. The ultimate goal is to match the observations to the model, which is done by tuning the model's parameters. Therefore, the goal of the inverse problem is to determine the best set of parameters.

Unfortunately, an inverse problem is, by definition, an ill posed problem:

- a solution may not exist
- a solution may not be unique
- a solution may not be stable
- errors in the observational record may be difficult to incorporate

Given this cautionary warning, in many cases, an inverse method is still the best approach to a problem. One may cast an inverse problem as an optimization problem, an attempt to minimize the error between the model output and the observations.

Inverse problems can be used for many questions in glaciology, such as using surface velocities to understand basal traction under ice streams or ice-shelf stiffness parameters; using borehole temperatures to infer a history of surface temperature; or inferring accumulation rates from internal layers.

Summary of lecture by Richard Hindmarsh -- Saturday, 12 June, 2010

Factors that determine ice flow include

1) Basal roughness

2) Rheology of subglacial sediment (till)

3) Basal hydrology

4) Ice rheology

1) Basal roughness. A rough bed slows glacier flow by forcing the ice to flow around the bumps, instead of slipping freely at the ice-bed interface. The degree of slowing due to a rough bed depends on the geometry of the bumps (essentially their length and height), and the number of contact surfaces between the ice and each bump. The original analysis of this problem (by Weertman) found that bumps of 1 m on a side were most effective in holding back glacier flow, assuming that the bumps are cubes.

2) Till rheology. If slippery sediment is present at the bed of an ice sheet or glacier, the overlying ice will move more quickly. The rheology of subglacial sediment is still debated; opinion varies as to whether this sediment behaves like a plastic material, or a viscous one. Squeezing till in the laboratory indicates a plastic rheology. However, the plastic rheology is inconsistent with the large volumes of sediment that are transported by glaciers. The viscous rheology is more consistent with observed sediment transport, but is not supported by lab experiments. Ultimately, the viscous rheology may be a parameterization of observed behavior, rather than a physically based treatment of subglacial sediment.

3) Basal hydrology. Water at the bed of an ice sheet may allow enhanced flow, either by "drowning" bumps (lifting the glacier off asperities), or by simple lubrication. However, a channel at the bed of an ice sheet will tend to close over time, as the overburden pressure causes ice to flow into the channel. Thus, channels must be kept open by frictional heating produced by water flow through the channels.

There is a conflict between large channels and distributed drainage at glacier beds. Distributed drainage networks contribute to enhanced ice flow, whereas large channels do not. However, large channels tend to "win," because high water discharge actually reduces the pressure within channels. RH notes that distributed networks have never been observed, only inferred.

The canonical view of drainage networks at glacier beds suggests a seasonal cycle. During the winter, little meltwater is produced, so there is no water-induced slip. In early spring, more meltwater is produced, and a distributed network forms at the glacier bed. The speed of the overlying ice increases in response to this lubrication. In summer, more meltwater is produced, and the distributed drainage network collapses to a channel. The basal sliding slows again.

4) Ice rheology. The stress-strain relationship for polycrystalline ice is usually represented with Glen's flow law, which has a temperature-dependent prefactor and an exponent n. Classic laboratory experiments suggest that Glen's n is 3, but these experiments all used strain rates that are two to three orders of magnitude faster than those observed in real ice sheets. That is, the time scale of a laboratory experiment is a few years, but the time scale of ice flow is hundreds to thousands of years. Other data sets suggest n values ranging from 2 to 4.5. An alternative flow law, the Goldsby-Kohlstedt flow law, is more consistent with the data, but produces results much like those of Glen's law in actual practice.