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INTRODUCTION TO PHYSICAL OCEANOGRAPHY

UNIVERSITY OF VICTORIA

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Foreward

These are the course notes for Descriptive Physical Oceanography, taught at the University of Victoria, School of Earth and Ocean Sciences, a 13-week undergraduate offering.

Unlike many physical oceanography courses, students taking this course are not expected to have vector calculus or differential equations, nor are they expected to have classical mechanics beyond what is typically offered in a first year physics course. Many of the students taking the course have a deep interest in understanding the ocean, and often approach the course with considerable intuition. The challenge of a course like this is to more than just a list of facts, but rather to try and give a feeling for the forces that drive the ocean, without becoming too bogged down in mathematical formalism. As such this course tries to strike a balance between knowing about the various physical phenomena in the ocean and understanding the dynamics that gives rise to them, while still giving the students some practice in practical calculations.

The course is divided into three unequal parts - overturning circulations driven by lateral buoyancy differences and mixing; waves and tides; and wind-driven circulations. We start with estuaries as a nice introduction to overturning circulations, but also because the students do work in Saanich Inlet, and understanding estuaries early is very helpful.

Part I

Overturning circulations

6

Climate Change and Sea level

There are numerous effects of climate change on the ocean, but perhaps two of the most fundamental are the effect of (all) melting ice on the thermohaline circulation, and the effect of melting land-fast ice and ocean warming on sea-level.

6.1 Effect of melting on the overturning circulation

North Atlantic Deep Water (NADW) is produced because it is salty and cold. As discussed in chapter 5, NADW sinks near Greenland and Iceland, and descends along the Western Boundary Current to depths of approximately 2500 m. Warm water from equatorward moves in to replace the sinking water. The overturning circulation in the North Atlantic is about 15-20 Sv and carries 1-1.5 PW of heat northwards [?].

If there is a change of the freshwater cycle or sudden introduction of fresh water in the northern hemisphere, say due to melting Arctic ice, or the Greenland ice sheet, then the overturning circulation can be suppressed, or stopped altogether (figure ??a, figure 6.1b). This stops the transport of heat northwards due to the overturning circulation, and has a dramatic effect on the temperature of the northern hemisphere, causing a cooling over the Atlantic in the simulation of up to 6 °C (figure ??c).

The role of the AMOC in warming the North Atlantic is relatively clear, but its role in sequestering anthropogenic CO₂ is less-well understood. However, it can readily be imagined that shutting off one of the sinking regions for surface water may indeed cause a decrease in the rate of ocean uptake of CO₂, leading to a deleterious feedback of warming the ocean further.

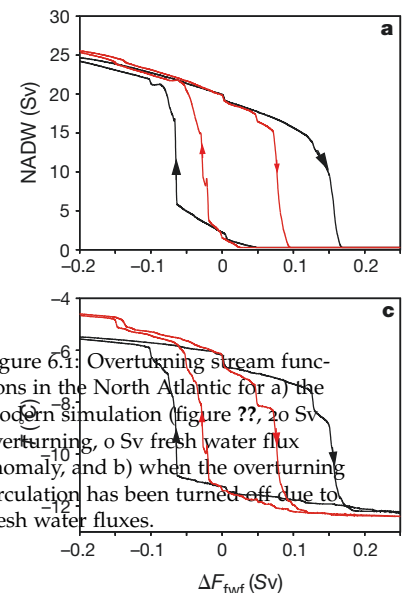


Figure 6.1: Overturning stream functions in the North Atlantic for a) the modern simulation (figure ??, 20 Sv overturning, 0 Sv fresh water flux anomaly, and b) when the overturning circulation has been turned off due to fresh water fluxes.

Figure 6.2: Effect of increasing fresh-water flux on a) the North Atlantic overturning circulation, and b) North Atlantic air temperatures 60-70 N ?. Red curves are fresh water introduced north of 50 N, black are south of 50 N. Note that restoring the fresh water flux to modern values does not allow the overturning circulation to immediately restart.

This remarkable re-arrangement of the circulation cannot persist without a large-scale change of the hydrological cycle, but it can introduce changes on decadal-to-century time scales. There is substantial evidence of rapid freshening of the North Atlantic during glacial times, in what are called **Heinrich Events**. These were originally discovered as open-ocean deposits of terrigenous rocks by large icebergs that broke off continental glaciers. While the chronology is not perfect, these events were also correlated with cooling, as inferred in Greenland ice cores. These coolings are known as **Dansgaard-Oeschger events**, and are largely confined to the northern hemisphere (

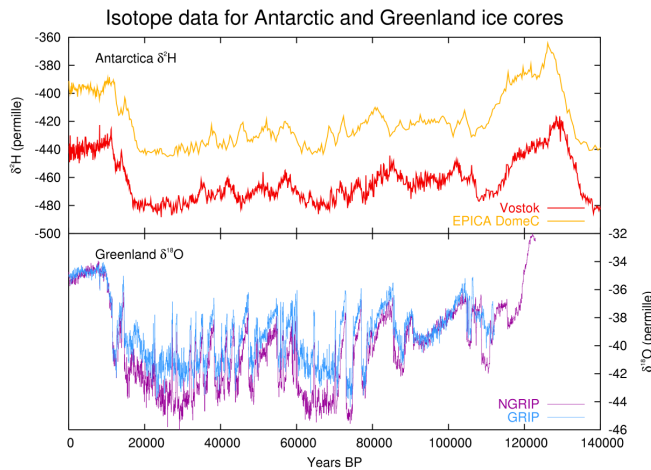


Figure 6.3: Ice core records of $\delta^{18}\text{O}$, a proxy for atmospheric temperature, with high values representing warmer atmosphere. (Leland McInnes at English Wikipedia <https://commons.wikimedia.org/wiki/File:Ice-core-isotope.png>)

Given the importance of the AMOC for our climate, there have been modern attempts to constrain its magnitude. These attempts actually contain a bit of a cautionary tale; considering 5 cruises from 1957 to 2004, [?] inferred a substantial slowdown of the transport across 24.5 N, from 23 to 15 Sv in the upper 1000 m (figure 6.4), and an accompanying slowdown of heat transport of approximately 1 PW.

Such a slow down was considered quite alarming and spurred better observations. Since 2004, the same line at 25 N was occupied, and supplemented by a mooring array to constrain the transport better. At the same time, estimates have been made at 35 S, 16 N, and approximately 55 N (figure 6.5). These demonstrate remarkable inter-annual variability, such that the result in [?] is likely a sampling error due to infrequent ship-based sampling. It is of note, however, that these four timeseries are not very well correlated, indicating substantial error and/or natural variability. There has been some argument that there has been slowdown in the RAPID array data, but much more subtle than that previously argued. Conversely, the array at

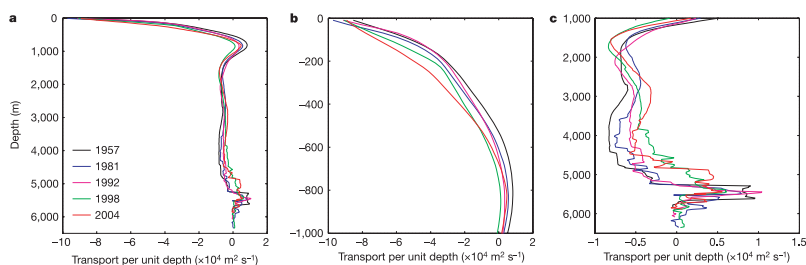


Figure 6.4: Transport as a function of depth across 24.5°N (approximately) [?].

16N indicates an increase in the overturning circulation. Continuing to monitor this overturning circulation is a large enterprise in global oceanography at present, with most Atlantic nations (including Canada) contributing.

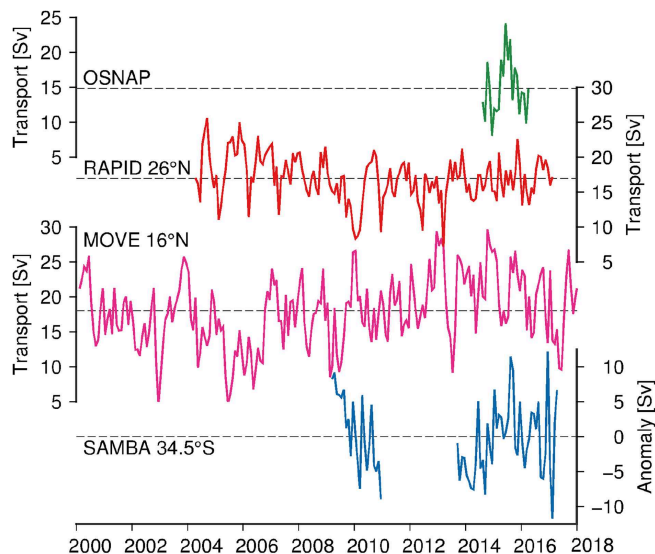


Figure 6.5: Northwards AMOC transport timeseries at 4 latitudes in the Pacific (OSNAP is approximately 55°N). Data gaps are due to arrays not being in place at those times.

6.2 Sea level change

A major physical effect on the ocean due to climate change is an increase in sea level, or ocean volume. This happens due to melting of land-fast ice (“Eustatic change”) and due to heating of the ocean water mass (“Steric change”). Since 1900, sealevel is believed to have increased by 20 cm. Over the last 20 years, melting has been estimated to be responsible for approximately 6 cm of sea level rise, whereas heating has been responsible for 3 cm.

GLOBAL SEA LEVEL

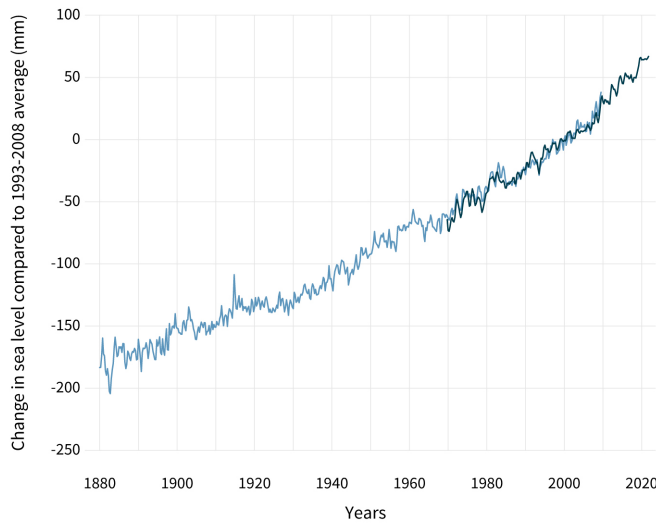


Figure 6.6: Global sea level change since 1880; blue is largely from tide gauges, black from satellite record (from <https://www.climate.gov/news-features/understanding-climate/climate-change-global-sea-level>, long timeseries is from [?]).

6.2.1 Land-fast ice

Melting of landfast ice is responsible for something close to $2/3$ of the sealevel rise observed in the last 20 years, and responsible for much of the change since 1900. Of course, floating ice is melting as well, but it displaces its own weight in water, so when it melts it does not add to the total volume of the oceans. Tracking where this water has come from is a major undertaking in glaciology.

The biggest advances in monitoring the landfast ice sheets have used satellite measurements using either lasers, radars, or gravimeters to determine changes in ice cover. These methods are in general agreement, and indicate that 4000 Gt of ice were lost from both ice sheets, with substantially more lost from Greenland than the Antarctic. Distributed over the area of the oceans this contributed 1 cm of sea level rise between 1990 and 2010.

6.2.2 Thermal expansion (steric change)

The other way that the ocean is expanding is due to thermal expansion as the ocean has warmed. The ocean has warmed about 2×10^{23} J between 1960 and today (figure 6.9). This means that, roughly, the ocean has increased its average temperature 0.03°C .

The warming effect is relatively easy to understand; suppose you have 1 L of 32 pus water at 20 degrees C - its density is $1022.504 \text{ kg m}^{-3}$. If we warm it 0.03 degrees, the new density is 1022.496. The mass will not have changed, so its new volume will be 1.0000078 L. If we multiply this over the whole ocean, we get a 3 cm of change over

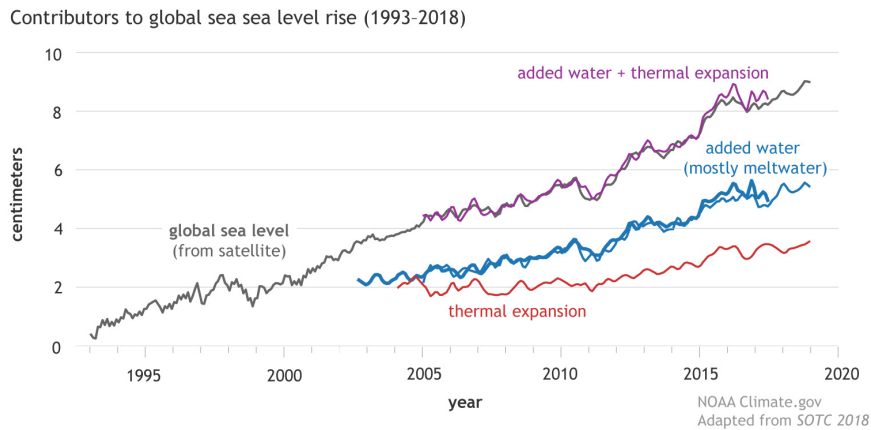


Figure 6.7: Contributions to sea level rise since 1993 (from <https://www.climate.gov/news-features/understanding-climate/climate-change-global-sea-level>).

4000 m. These rough numbers are similar to what is shown in figure 6.7.

This heat change is difficult to measure over the whole globe, and requires integrating temperature estimates from many measurements. Ship measurements get to the deepest depths, and then the more variable surface waters are more comprehensively measured with a fleet of autonomous profiling floats called Argo floats

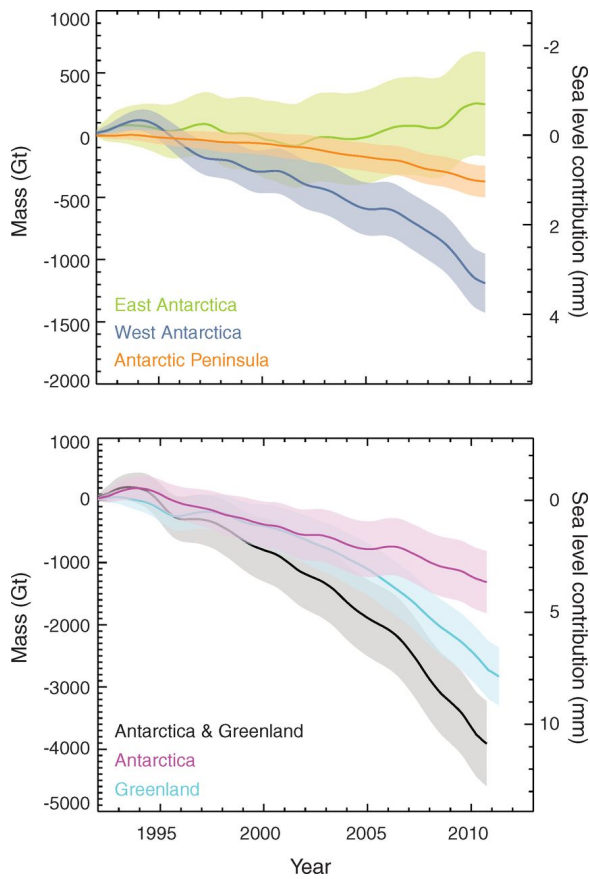


Figure 6.8: a) Cumulative ice sheet mass loss over time in regions of Antarctica, b) all Antarctica and Greenland [?].

OCEAN HEAT COMPARED TO AVERAGE

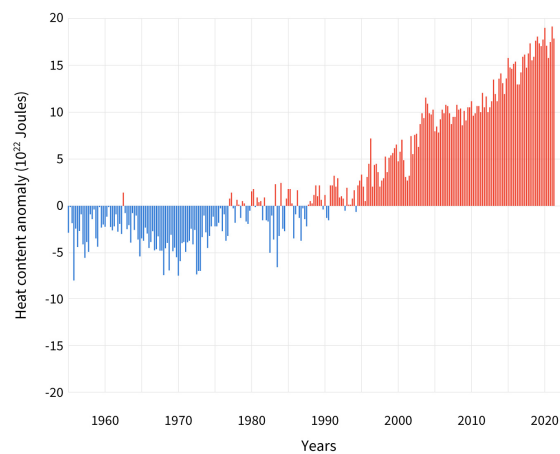


Figure 6.9: From <https://www.climate.gov/news-features/understanding-climate/climate-change-ocean-heat-content>

Part II

Waves and Tides

Part III

Wind-driven circulations