Engineering 4 (2018) 6–8

Contents lists available at ScienceDirect

Engineering

journal homepage: www.elsevier.com/locate/eng

Views & Comments

Glaciology and Global Climate Change

John C. Moore [a,b]

[a] College of Global Change and Earth System Science, Beijing Normal University, Beijing 100875, China
[b] Arctic Center, University of Lapland, Rovaniemi 96100, Finland

If it had been suggested prior to the 1980s that the study of ice—that is, glaciology—would be the key to initiating international action on global climate change, most climate scientists would probably have laughed. However, that is exactly what occurred. The kick-start came from the Vostok ice core, a project that had been underway at the Soviet station at the Pole of Inaccessibility in East Antarctica since the 1970s. By the mid-1980s, the French were in close partnership with the Soviets; in 1987, they jointly published their results on the stable water isotopes and greenhouse gas concentrations (methane and carbon dioxide, CH_4 and CO_2) identified in air bubbles in the core. The implications were immediately obvious to all: There was a clear correlation between the concentrations of the greenhouse gases and the air temperatures over the ice sheet (revealed by the isotopic ratios in the ice), as shown in Fig. 1 [1,2]. The steady beat of the glacial-interglacial cycle over some 110 000 years could be seen in the 2 km deep ice core record. Although the temperature signal was responding to changes in the Earth's orbital geometry, as shown previously in deep sea sediment records, greenhouse gases were playing an amplifying role in the behavior of the temperature signal. In the following year, 1988, the United Nations launched the Intergovernmental Panel on Climate Change (IPCC), leading to the regular IPCC reports that summarize the current state of knowledge regarding the climate system.

The paleoclimate record that has been obtained from ice cores drilled in Antarctica and Greenland represents global climate variability on timescales of millennia and longer. Faster fluctuations are affected by the local peculiarities of Antarctica and Greenland, leading to interesting differences in the ice core records [3]. These differences are fundamentally caused by two factors. First, Greenland is located south of the Arctic Ocean, and many other landmasses are located almost as far north as Greenland. Eastern Canada and Scandinavia were occupied by large ice sheets during the glacial periods, which dominated the last few hundred thousand years, whereas Greenland is the only Northern Hemisphere ice sheet existing during the relatively short (10 000 year) interglacial periods. The 3 km high Laurentide Ice Sheet and Scandinavian Ice Sheet produced different atmospheric circulation patterns than those seen today. In contrast, Antarctica is surrounded by the Southern Ocean, along with other significant landmasses that are located too close to the Equator to support large ice sheets, even during the cold glacial periods; therefore, the main circulation patterns remain largely unaltered over glacial cycles. The second reason why Greenland reacts differently than Antarctica is the role of the North Atlantic Ocean in circulating warm tropical ocean waters to the north and into the Arctic seas. At present, these currents provide about 30% of the heat to Northern Europe, making that region much warmer than corresponding latitudes in Asia. This system (known as the Atlantic meridional overturning circulation, AMOC) appears to be rather sensitive to changes in the density and temperature structure of the ocean. During much of the glacial period, this circulation was shifted far to the south as compared with today; however, it could occasionally move to similar locations as at present, dramatically altering local temperatures, sea ice extent, moisture supply to the Greenland Ice Sheet, and so forth. This AMOC variability makes the Greenland paleoclimate record far more variable than that recorded in Antarctica. Climate models suggest that under increased greenhouse gas concentrations, the AMOC will steadily decrease over the next century or two [4].

If ice core science showed that the climate of the past was both remarkably and unexpectedly variable, then the visual evidence of glacier change since the end of the Little Ice Age (around the mid-19th century), seen in paintings and later in photography, shows dramatic evidence of ongoing climate change. Glaciers in mountain regions across the world reached their maximal extents in the 19th or early 20th centuries, thus providing unequivocal evidence of the changes that are occurring on century timescales, even to these apparently immobile rivers of ice. However, all the ice contained in the 200 000 or so mountain glaciers around the world would only raise the global mean sea level by about 40 cm by melting [5]. Although this would be an important change, it would not yield the catastrophic levels of sea level rise that would question the resilience of coastal cities and infrastructure. But if the Greenland and Antarctic ice sheets were also to react to warming on a century timescale, then their potential to raise global sea level is overwhelming: Enough water to raise the sea level by 7 m is stored in Greenland, and enough to raise it by 65 m is stored in Antarctica.

The important question challenging many ice modelers today is, how much sea level rise can we expect to come from the ice sheets over the next century or two? This question was brought into focus by the recent dramatic breakup of many large ice shelves around the fringes of Antarctica [6], and by the suggestion that past sea level rise sensitivity and response speed to warming was far faster

https://doi.org/10.1016/j.eng.2018.01.001

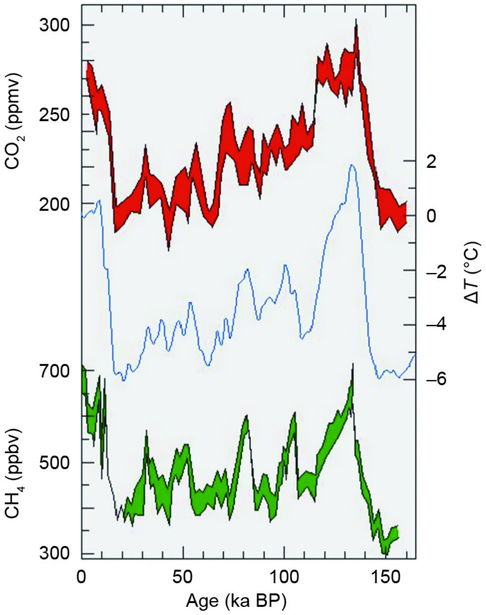

Fig. 1. 160 000 years of air temperature (blue line) and greenhouse gases (red line for CO_2 and green line for CH_4) recorded in the Vostok ice core [1,2]. ppmv: parts per million by volume; ppbv: parts per billion by volume; ka BP: 1000 years before the present.

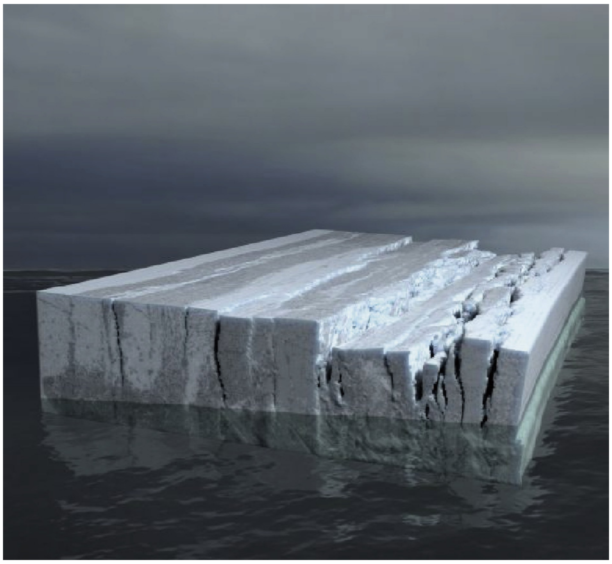

Fig. 3. Discrete particle model simulation of a partially submerged ice block that is 80 m high and 500 m long. The ice block terminates in 40 m deep water, fracturing under its own weight. The individual particles are 1 m^3 in size, and are designed to mimic the visco-elastic behavior of ice. Note the realistic distribution of fragment sizes and the size of fractures, which are quite different from those of a simple rock or sand pile avalanche. (Graphic rendered by Jyrki Hokkanen, CSC–IT Centre for Science Ltd.)

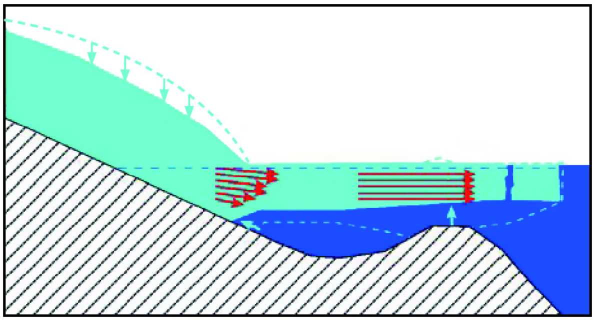

Fig. 2. Cartoon of the dynamic impact of ice shelf loss on land-based glaciers due to the removal of the buttressing effect. The dashed green profile shows how the glacier and ice shelf profile would be in the case of the ice shelf providing significant backstress via the bedrock high beneath the floating ice shelf. If the ice shelf thins, as shown by the solid-colored profile, typically via bottom melting (green upwards arrow), then the ice shelf loses contact with bedrock highs, reducing buttressing force. This allows ice to flow faster off the land (red arrows), lowering the inland glacier (green downward arrows) contributing to sea level rise. Once the ice shelf loses contact with the land, it typically breaks up rapidly and retreats inland (illustrated by the fracture near the ice shelf terminus). The blue dashed line shows the sea level.

than anticipated for these huge inertial systems [7]. The ice shelves—that is, floating glacier ice, not frozen sea ice—that form as glaciers flow off the continental ice sheet into the sea and act as dams against the free flow of inland ice. The ice shelves exert a buttressing force by grounding on the shallows of the continental shelf sea floor (Fig. 2).

Ice shelves can disintegrate very rapidly; in just a few years or decades, thousands of square kilometers of thousand-year-old ice may break up. There are two main mechanisms for this breakup [6]. In the first process, surface meltwater created by warm air temperatures fills the bottom of crevasses or depressions on the ice shelf. This meltwater absorbs solar radiation and warms up due to its low albedo; eventually, it causes the ice shelf to fracture all the way to its base, even hundreds of meters below the surface, in a process known as “hydro-fracturing.” The second process is via basal melting and weakening by warm ocean water, which can typically deliver an order of magnitude more heat to the ice shelf than the atmosphere does. As the ice shelf thins, it becomes unable to support its own weight, and simply breaks apart [8] (Fig. 3), behaving as a self-organized critical system.

Rising temperatures promote both methods of rapid ice shelf decay and the removal of buttressing force. But to allow fast flow of the inland ice, that ice needs to be on reverse-sloping bedrock; that is, the bed must get deeper inland. This is the case for many of the large glaciers filling troughs in the Amundsen Sea and Bellingshausen Sea sector of West Antarctica.

With only a decade of data, the GRACE (short for Gravity Recovery and Climate Experiment) gravity anomaly satellite already shows that this region, which encompasses the Pine Island and Thwaites glaciers, is quite significantly in negative mass balance [9]; that is, mass loss from the ice sheet to the ocean is already occurring. Future rates of mass loss depend on critical details of bedrock geometry and on the bathymetry of the ocean cavity that connects to the Southern Ocean. For rapid melting to occur, warm water must be able to access the ice shelf cavity; this warm water is typically quite deep (500 m or more), since it is much denser than the lighter, cooler surface waters. However, a typical ice shelf begins to float in water depths of a kilometer or so. This grounding line is where melting occurs fastest, and where the critical transition occurs from ice that is slowed down by friction with bedrock, to a freely floating ice shelf.

Thus, fruitful areas of research in the field of global sea level rise include: fracture processes in ice dynamics models that are designed for fluids of very high-viscosity [8], and ice shelf-ocean interaction [10]. Although these topics are rather immature, their results are required by planners, engineers, and politicians who need to know how much local sea levels may rise over the coming century, and especially how large sea level extreme flood events may be [11].

Acknowledgements

Research funded by the National Basic Research Program of China (2015CB953602).

References

[1] Barnola JM, Raynaud D, Korotkevich YS, Lorius C. Vostok ice core provides 160,000-year record of atmospheric CO_2. Nature 1987;329(6138):408–14.

[2] Lorius C, Jouzel J, Raynaud D. The ice core record: Past archive of the climate and signpost to the future. Phil Trans R Soc Lond B 1992;338(1285):227–34.

[3] Barker S, Knorr G. Antarctic climate signature in the Greenland ice core record. Proc Natl Acad Sci USA 2007;104(44):17278–82.

[4] Cheng W, Chiang JCH, Zhang D. Atlantic meridional overturning circulation (AMOC) in CMIP5 models: RCP and historical simulations. J Clim 2013;26(18):7187–97.

[5] Moore JC, Grinsted A, Zwinger T, Jevrejeva S. Semiempirical and process-based global sea level projections. Rev Geophys 2013;51(3):484–522.

[6] Liu Y, Moore JC, Cheng X, Gladstone RM, Bassis JN, Liu H, et al. Ocean-driven thinning enhances iceberg calving and retreat of Antarctic ice shelves. Proc Natl Acad Sci USA 2015;112(11):3263–8.

[7] Grinsted A, Moore JC, Jevrejeva S. Reconstructing sea level from paleo and projected temperatures 200 to 2100 AD. Clim Dyn 2010;34(4):461–72.

[8] Åström JA, Vallot D, Schäfer M, Welty EZ, O'Neel S, Bartholomaus TC, et al. Termini of calving glaciers as self-organized critical systems. Nat Geosci 2014;7(12):874–8.

[9] King MA, Bingham RJ, Moore P, Whitehouse PL, Bentley MJ, Milne GA. Lower satellite-gravimetry estimates of Antarctic sea-level contribution. Nature 2012;491(7425):586–9.

[10] Asay-Davis XS, Cornford SL, Durand G, Galton-Fenzi BK, Gladstone RM, Gudmundsson GH, et al. Experimental design for three interrelated marine ice sheet and ocean model intercomparison projects: MISMIP v.3 (MISMIP+), ISOMIP v.2 (ISOMIP+) and MISOMIP v.1 (MISOMIP1). Geosci Model Dev 2016;9(7):2471–97.

[11] Jevrejeva S, Grinsted A, Moore JC. Upper limit for sea level projections by 2100. Environ Res Lett 2014;9(10):104008.