Irregular oscillations of the West Antarctic ice sheet

Douglas R. MacAyeal

Department of Geophysical Sciences, University of Chicago, 5734 South Ellis Avenue, Chicago, Illinois 60637, USA

Model simulations of the West Antarctic ice sheet suggest that sporadic, perhaps chaotic, collapse (complete mobilization) of the ice sheet occurred throughout the past one million years. The irregular behaviour is due to the slow equilibration time of the distribution of basal till, which lubricates ice-sheet motion. This nonlinear response means that predictions of future collapse of the ice sheet in response to global warming must take into account its past history, and in particular whether the present basal till distribution predisposes the ice sheet towards rapid change.

CONCERN that the West Antarctic ice sheet (WAIS) might collapse in response to future climate warming is motivated by geological evidence suggesting its collapse some time during the past million years. (Here I use the term collapse to indicate the conversion by ice flow of a partially grounded, partially floating ice sheet into a purely floating ice shelf in such a manner as to increase the global ocean volume.) The evidence traditionally cited for possible WAIS collapse involves the interpretation of fossil beach and coral-reef terraces. Some interpretations suggest that sea level was 6 m higher during the 120 kyr BP (before present) interglacial (Sangamon, or isotopic stage 5e) than it is today[1]. If the Greenland and East Antarctic ice sheets were roughly the same size during the Sangamon as they are today (an assertion subject to question[2]), such a high sea-level stand would point to a collapse of the WAIS[3]. This interpretation may be in doubt because the 6 m sea-level high-stand could actually represent a geodynamic effect associated with glacial rebound[4] rather than a change in ocean volume. Nevertheless, support for possible WAIS collapse comes from the West Antarctic region itself: U/Th chronology of carbonate lacustrine deposits in the dry valleys[5] (along the western margin of WAIS) confirms that there was a marked retreat of grounded ice between 130 and 98 kyr BP. Moreover, marine microfossils of age less than one million years are found in glacial till below grounded portions of the WAIS[6]. This suggests, but may not establish (L. H. Burckle, personal communication), seawater invasion at least once during the past million years.

Given concern over the future of the WAIS, and the likelihood that the geological record alone will not reveal the exact timing or cause of ice-sheet collapse, I use a numerical model of ice-sheet, ice-stream and ice-shelf dynamics to determine what attributes of dynamics and climate forcing could have caused WAIS collapse during the Late Pleistocene. I shall demonstrate that the long-term climate history of Antarctica associated with the glacial and interglacial cycle could have caused the WAIS to collapse and re-form several times over the past million years. The climate history is revealed by the surface temperature record derived from the Vostok ice core[7], and the history of global sea level is estimated from isotopic records in deep-sea sediments[8] (Fig. 1). The special characteristics of ice streams, most notably the dynamics introduced by a deformable bed, suggest a geological history that can include sporadic, perhaps chaotic, cycles of collapse and re-formation. The reason for this behaviour is that the timescale needed to equilibrate the distribution of deformable subglacial till is comparable to the 100,000-year timescale of the glacial cycle. The ice sheet thus cannot respond quasi-statically to long-term external climate forcing. Instead, it is jarred into possibly chaotic behaviour by the external forcing, and collapses occasionally when deformable basal till becomes widespread.

A whole ice-sheet model

To investigate the history of WAIS during the past million years, I constructed a finite-element model of ice-sheet flow and mass balance that can reproduce the present-day flow regime of the ice sheet, including its ice streams and ice shelves. This model represents a set of conservation rules which determine the time evolution of ice thickness in response to snow accumulation, basal melting, iceberg discharge, the horizontal divergence of ice flow dictated by gravitational driving stresses, and the temperature of the ice and its bed. Also included are conservation rules regarding the thickness and thermodynamic state of

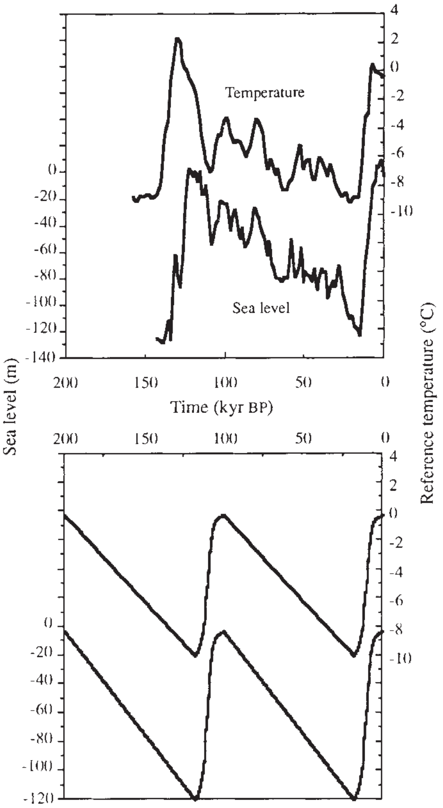

FIG. 1 Upper panel, Antarctic surface-temperature change chronology (upper curve) derived from the Vostok ice core[7] and a global sea-level chronology (lower curve) estimated from deep-sea sediment isotope records[8]. Lower panel, abstract chronologies of surface temperature change (upper curve) and global sea level (lower curve) for the simple sawtoothed climate cycles used as forcing in the model simulation discussed in Fig. 3.

deformable subglacial sediments, which are the hallmark of WAIS ice streams[9].

Inland-ice dynamics similar to those used successfully in previous studies of Antarctic glaciation[10] determine the ice flow and mass balance where the ice sheet is frozen to its bed (basal sliding is not allowed). Ice-stream dynamics, including a treatment of the mechanics and thermodynamics of deformable subglacial till, are used where the ice-sheet bed reaches the melting temperature, and in elements that bridge the transition zones between wet and frozen bed[11]. Ice-shelf dynamics are used where the ice sheet floats free of the subglacial bed[12]. A heat-flow equation accounting for vertical conduction and advection predicts the temperature–depth profile in the ice. In the upper 2,000 m of the subglacial bed, this profile is modelled using a vertical heat conduction equation and a geothermal flux basal boundary condition ($0.055\ \mathrm{W\,m^{-2}}$)[13] to account for long-term thermal inertia effects. Horizontal heat diffusion and advection are disregarded because they cannot be treated accurately at the spatial resolution of this study. Latent heat and strain-heating effects in subglacial till are accounted for when determining the melting rate at the base of the ice. Crustal deformation in response to changing ice and seawater loads is treated in a simplified manner using an e-folding relaxation time constant of 8,000 years[14]. Elastic flexure and spatial length-scale effects[15] are disregarded for simplicity.

The processes represented by this model are simplifications of those required for more comprehensive treatments of ice flow in West Antarctica. In my opinion, a simpler model could not capture the essential behaviour of the WAIS and a more complex model would be computationally unwieldy for million-year-long simulations (taking 25-year time steps). Work in which the model, or pieces of the model, has explained observations of the present WAIS[10,11,16,17] defends my simplifications.

A critical feature of the model, and one in which it differs from previous models used to investigate WAIS, is its treatment of deformable basal sediments, the lubricant of ice streams. Relatively little is known about the mechanical and hydrological properties of deforming till layers beneath West Antarctic ice streams. Field observations suggest that they are ~5–10 m thick, are dilated and filled with pore water near the overburden pressure, and are so inviscid compared with ice that most of the horizontal ice velocity is accommodated by shear within the till[18,19,26]. I account for four of the most important qualitative aspects of deformable bed physics. First, deforming till satisfies a mass-continuity equation (in which vertically integrated horizontal till-flux divergence is balanced by till-layer thinning and the creation or removal of deforming till at the till/rigid-sediment interface below the deforming till layer). Second, water is required to fill void space in newly dilated till (or is expelled when till ceases to deform and begins to consolidate). Third, strain heating is localized in till, and latent heat (of water in the pore space) is stored in till. Finally, till affects the basal friction of the ice sheet.

I use a vertically averaged continuity equation that treats horizontal till-flux divergence and erosion of the substratum to predict till-layer thickness. To compute till-flux, the vertically averaged horizontal velocity is assumed to be half the ice velocity at the ice/till interface[20]. This assumption is correct if the till is a viscous fluid. For other rheological properties[19], the vertically averaged till velocity would be less than the value assumed here. The conversion of rigid, consolidated substratum into the mixture of unconsolidated sediment and water that comprises deformable till is assumed to be limited by the availability of water. A rationale for this assumption is that water must be drawn through the relatively impermeable till to the till/rigid-substratum interface where void space is created. The pore-water pressure gradient necessary to induce this water flux exerts an important influence on the conversion rate[21]. The conversion rate is thus expected to be a function of the local basal-melting rate and the rate at which basal meltwater is dissipated by subglacial drainage mechanisms. I assume a linear function. (More advanced models would require explicit treatments of the basal water conduit system, the pore-water pressure field, the basal stress imposed by the ice sheet, and the tenacity of the substratum.) If the basal melting rate is positive, the constant of proportionality is equal to $(1-f)/\theta$, where f is the fraction of total basal meltwater production dissipated by subglacial drainage and θ is the porosity difference between rigid substratum and till (θ is expressed as a non-dimensional ratio of pore volume to solid volume). If the basal melting rate is negative (freezing), I assume that subglacial drainage is shut down and the constant of proportionality is thus $1/\theta$. The value of f is not well constrained by observations, but the porosity of deforming till below ice stream B is 0.4 (ref. 22). To yield computed till thickness which falls in the observed range of 5–10 m, I take f and θ to be 0.75 and 0.4, respectively.

The effect of deformable basal till on the stress balance of the ice sheet is uncertain because of widely different interpretations of available data. I adopt the 'sticky-spot' interpretation, which holds that subglacial till is so inviscid that the large-scale basal friction is determined by isolated bedrock pinnacles that poke through the till layer, or by other lithological features of the glacial substratum[19]. Observations of basal-stress distributions[17] suggest that sticky spots have small spatial scales (1–10 km) which fall below the finite-element resolution of the model used here. I therefore use a spatially uniform, empirically determined[17] basal-friction coefficient ($1.5\times10^9\ \mathrm{Pa\,s\,m^{-1}}$) in the ice-stream stress-balance calculation to represent the average large-scale effect of sticky spots. I define the basal-friction coefficient as a scalar that, when multiplied by basal ice velocity, gives the direction and strength of the basal traction. Although it is premature to accept the sticky-spot interpretation over other competing theories, I use it because it is simple, has been successful in simulating the observed velocity field of ice stream E (ref. 17) and gives the WAIS model the ability to produce till in amounts currently observed and to produce ice streams that move with the correct velocity scale. Sensitivity tests not reported here suggest that plausible alternative basal-till rheological

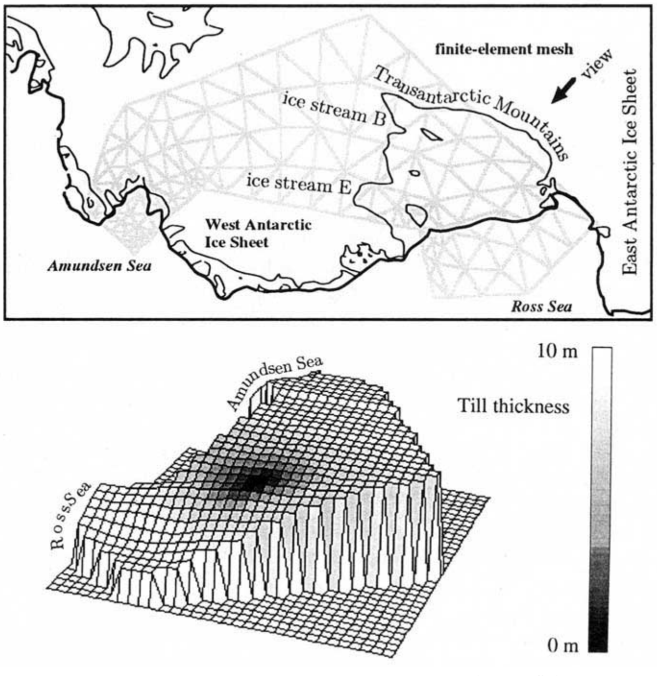

FIG. 2 Upper panel, west Antarctica and its low-resolution finite-element representation. Lower panel, an oblique view (taken from a viewpoint west of the Transantarctic Mountains) of the WAIS model surface elevation during typical interglacial conditions. Shading on the surface of the ice sheet represents the thickness of deformable subglacial till which lubricates the bed. Maximum till thickness in this particular configuration of WAIS (~10 m) occurs in roughly the area in which ice streams are found today.

properties, such as those of a purely viscous fluid[20], produce ice-sheet behaviour that is even more unstable than that described below.

Having simplified its physical processes, I also reduce the complexity of WAIS geometry to make the computation of WAIS evolution over long time periods manageable. Accordingly, I take the curved channel extending from the open Ross Sea to the Amundsen Sea through the region of present ice-stream activity (Fig. 2) to be the computational domain of WAIS. Most notably absent in the simplified geometry are the narrow ice streams which drain between interstream ridges along the boundary between inland ice and ice shelf. I forego the geometric but not the dynamical accuracy of these ice streams. As stated above, regions determined to rest on a melted bed are allowed to evolve according to the stress- and mass-balance equations that have successfully simulated ice streams in previous high-resolution models[11,17]. Two boundaries of the idealized WAIS model domain separate WAIS from East Antarctica and from the relatively small but stable ice dome that sits on the coast between the Amundsen and Ross Seas. These boundaries are assumed to be rigid and impermeable; ice flow parallel and perpendicular to these walls is therefore zero in the model. The two boundaries that face the Ross and Amundsen Seas are treated as open, so that they freely allow the calving of icebergs into the ocean. For the stress balance at the open boundaries, I specify seawater pressure as the dynamic boundary condition. The undisturbed elevation of the continental shelf below the idealized WAIS geometry is assumed spatially uniform at a value of 350 m below sea level when ice is absent, isostatic rebound is complete, and global sea level is at its present value.

Climatic forcing is represented by two agents: local surface temperature (which affects local precipitation) and global sea level (which affects ice-sheet grounding lines). Following previous ice-sheet modelling conventions, ice-sheet surface temperature is taken as a function of ice-sheet elevation, an atmospheric lapse rate, and a reference sea-level air temperature which is assumed to vary in time according to the Vostok ice-core record of temperature change (Fig. 1). I prescribe a standard lapse rate for atmospheric temperature based on empirical data ($5\,^{\circ}\mathrm{C\,km^{-1}}$ below 1,500 m and $14\,^{\circ}\mathrm{C\,km^{-1}}$ above)[22]. This lapse rate has become a standard attribute of local Antarctic climate which allows intercomparison of Antarctic ice-sheet modelling studies. The reference sea-level air temperature (−22.5 °C) is assumed to be spatially uniform, but is allowed to vary in time by up to 12 °C as suggested by the Vostok ice-core temperature record[7]. Following previous ice-sheet modelling conventions, surface accumulation is specified to be the sum of a spatially uniform value of $0.25\,\mathrm{m\,yr^{-1}}$ (ice equivalent) and deviation which is a function of the local air temperature at the ice-sheet surface[7,13]. As with the lapse rate, this deviation function represents a standard attribute that allows model intercomparison.

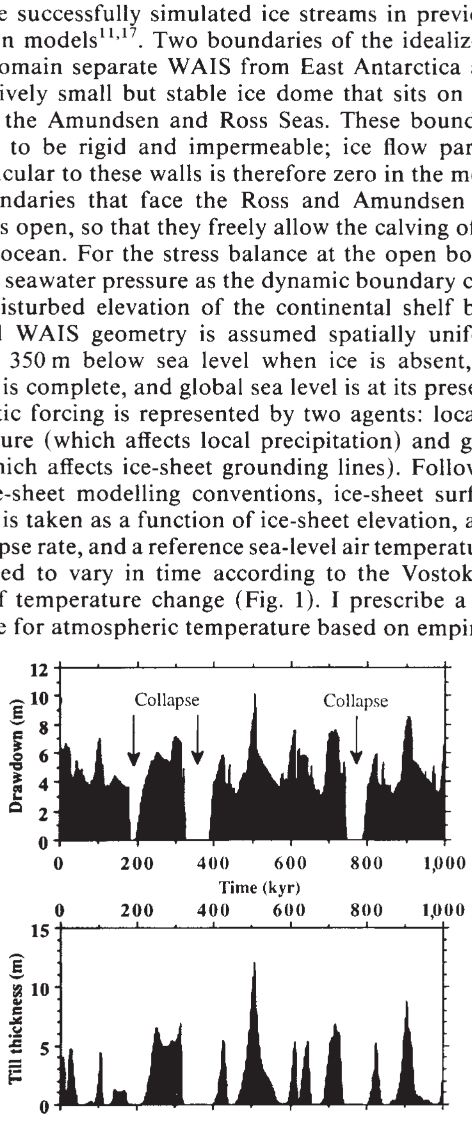

FIG. 3 Sea-level drawdown (upper panel) and average till thickness (lower panel) for a million-year simulation forced by the simple 100,000-year sawtoothed temperature and sea-level cycles in Fig. 1. Sea-level drawdown is a way of measuring the size of the WAIS. It represents the eustatic sea-level drop (in metres) required to build the particular configuration of the WAIS. This measure of ice-sheet size is zero when no grounded ice exists in West Antarctica and crustal rebound is complete. Ice-sheet collapse is thus signified when the sea-level drawdown is zero. The feedback effect of WAIS on sea level is not accounted for in the imposed external forcing of the ice sheet. Subglacial till thickness, averaged over the wet-bed fraction of the total ice-sheet footprint, is shown to verify model performance and to highlight the relation between till build-up and ice-sheet collapse. Typical till thickness observed below the West Antarctic ice streams today is 5–10 m (refs 9, 26).

Global sea-level forcing is prescribed by changing the elevation of the continental shelf on which the ice-sheet rests. This forcing is assumed to be caused by the advance and retreat of the Northern Hemisphere ice sheets. Effects of the WAIS on global sea level are disregarded because they are small (5% of the typical glacial to interglacial sea-level change).

Response to 100-kyr climate cycles

To determine what attributes of climate forcing, if any, could account for occasional collapse of WAIS, I ran a number of million-year-long simulations. Million-year runs were necessary to dissipate the influence of an arbitrary initial condition and to allow an ensemble of ten 100,000-year glacial cycles to be generated by each experiment for statistical analysis. Longer runs were prohibitively expensive. I prescribed cyclic surface-temperature forcing, cyclic global sea-level forcing, or both. The initial condition in all experiments was the same: a model-generated steady-state ice-sheet configuration associated with glacial atmospheric and sea-level conditions (reference sea-level air temperature 10 °C colder, and global sea level 120 m lower, than today).

I found that many combinations of surface-temperature and sea-level forcing could produce irregular ice-sheet fluctuations and sporadic collapse. This finding is best illustrated in Fig. 3 which shows the results of a simulation in which both the reference sea-level air temperature and global sea level were specified as the cyclic repetition of the simple sawtoothed 100,000-year time series shown in Fig. 1. Despite the simplicity and regularity of the climate forcing, the response of the ice-sheet is irregular. Three collapses occur (identified in Fig. 3 by the sea-level drawdown, a measure of ice-sheet size, going to zero) at ~190,000 years, ~330,000 years and ~750,000 years after the start of the simulation.

Response to transient change

To understand why the simulated history of WAIS discussed above is so irregular, it is constructive to examine the transient

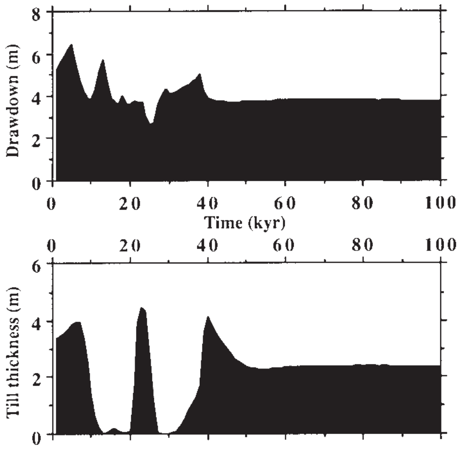

FIG. 4 Sea-level drawdown (upper panel) and average till thickness (lower panel) when the WAIS model is forced by a sudden switch between glacial and interglacial climate conditions.

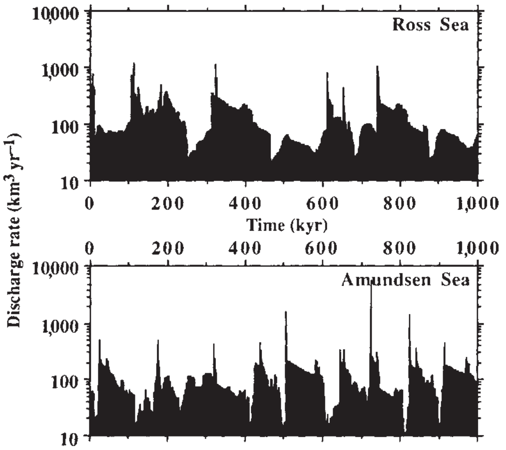

FIG. 5 Iceberg-discharge rates into the Ross (upper panel) and Amundsen Seas (lower panel) predicted by the million-year simulation of WAIS response to the simple sawtooth climate forcing shown in Fig. 1 (see also Fig. 3).

response to a sudden switch in climate from glacial to interglacial conditions. I will show that the timescale of this transient response is long enough to suggest that irregular behaviour described above is a simple consequence of mismatch between forcing timescale and the natural timescale of ice sheets which have deformable beds.

I simulated the ice sheet's evolution from an initial steady state (consistent with glacial conditions) used in the simulations discussed above towards a new steady state consistent with interglacial conditions (global sea level increased by 120 m and surface temperature increased by 10 °C over the glacial values). The WAIS was able to reach a stable steady state in this simulation, but only after a long period of adjustment. It took 50,000 years for the WAIS to equilibrate, and during this period there were several distinct oscillations of growth followed by rapid shrinkage (Fig. 4).

As seen by comparing the time series of average till thickness and the sea-level drawdown in Fig. 4, the slow build-up of deforming subglacial till towards a critical value seems to trigger rapid ice-sheet shrinkage. This is because ice-stream activity is generally dormant until the region containing deformable till connects with an open boundary through a frozen-bed zone that rings the margin of the ice sheet. Once this connection is established, ice-stream drainage into the Ross and Amundsen seas rapidly shrinks the ice sheet below what would eventually become its equilibrium size. Long timescales are introduced into this process because slow thermodynamic processes are involved in bringing frozen-bed zones up to the melting temperature.

The long timescale of WAIS adjustment to impulsive climate change (50 kyr) is comparable to the timescales of late-Pleistocene climate forcing (20, 40 and 100 kyr). In this circumstance, trends in WAIS configuration at any one instant in the past may not be in balance with the instantaneous state of external climate forcing.

Predictions

The simulations of WAIS evolution in response to a cyclic climate forcing suggest that the late Pleistocene history of the WAIS should be irregular. Verification of this irregularity may be possible by examining the geologic record for signs of sporadic collapse and periods of extreme iceberg discharge. Marine microfossil observations are consistent with the notion that collapse is a rare phenomena, and is not a periodic feature of the glacial cycle (ref. 6, and L. H. Buckle, personal communication). Figure 5 displays the iceberg discharge rates into the Ross and Amundsen seas associated with the million-year simulation discussed above (Fig. 3). Short periods of extreme iceberg discharge with amplitudes in excess of 500 km^3 yr^{-1} occur irregularly throughout the simulation. The occurrence of such extreme events may be detectable as prominent peaks in the ice-rafted debris abundance in offshore sediments, and perhaps as oxygen-isotope enrichments in local marine microfossils. I propose that the iceberg-discharge spikes simulated here may be West Antarctic equivalents of the Heinrich events thought to portray sudden, episodic iceberg discharge into the North Atlantic[24].

Another qualitative prediction gained from the simulations is the maximum rate of sea-level rise in response to WAIS collapse. The maximum rate for the simulation shown in Fig. 3 was ~0.25 m per century. Previous work using more realistic sub-ice topography and stronger climate forcing, but less-realistic ice-sheet physics, has suggested that larger rates of sea-level rise could be possible[25].

Conclusions

Model experiments indicate that climatic forcing associated with the glacial cycle could have elicited sporadic collapse and regrowth in the history of the WAIS. The irregularity of these events is a principal qualitative result which should be tested observationally by inspecting the marine record for evidence of sudden, strong iceberg discharges. On the prospect of forecasting WAIS response to CO_2-induced warming, I offer three opinions based on these experiments. First, collapse is not necessarily correlated with extremely warm or extremely cold periods of climate forcing. It is triggered by deformable till conditions at the bed of the ice sheet which, because it is isolated from the atmosphere, are effected by long-term climate trends (thousands of years) as well as the vagaries of ice-sheet history. Second, very long timescales (~50,000 years) can be involved in WAIS adjustment to simple, impulsive climate change because of the nonlinear nature of ice-stream dynamics. This suggests that forecasts of future behaviour will be sensitive to errors in the specification of (present) initial conditions which determine the ice sheet's predisposition towards rapid change. Third, the hypothesis that WAIS snow-accumulation rates increase when the climate warms does not contradict the hypothesis that future climate warming could cause the WAIS to collapse. My simulations incorporate the feedback between accumulation rate and climate warming, yet they display ice-sheet collapse or shrinkage at times when model climate forcing is warm. What is most important in determining the possibility for WAIS collapse in the near future is the distribution of deformable subglacial till.

Received 29 April; accepted 7 July 1992.

1. Chappell, J. & Shackleton, N. J. *Nature* **324,** 137–140 (1986).
2. Letreguilly, A. *et al. J. Glaciol.* **37,** 149–157 (1991).
3. Hughes, T. *Rev. Geophys. Space Phys.* **13,** 502–526 (1975).
4. Lambeck, K. & Nakada, M. *Nature* **357,** 125–128 (1992).
5. Denton, G. H. *et al. Quat. Res.* **31,** 151–182 (1989).
6. Scherer, R. P. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **90,** 395–412 (1991).
7. Lorius, C., Raisbeck, G., Jouzel, J. & Raynaud, D. *The Environmental Record in Glaciers and Ice Sheets* (eds Oeschger, H. & Langway, C. C. Jr) 343–361 (Wiley, Chichester, 1989).
8. Shackleton, N. *J. Quat. Sci. Rev.* **6,** 183–190 (1987).
9. Alley, R. B. *et al. Nature* **322,** 57–59 (1986).
10. Huybrechts, P. *Clim. Dynam.* **5,** 79–92 (1990).
11. MacAyeal, D. R. *J. geophys. Res.* **94,** 4071–4087 (1989).
12. MacAyeal, D. R. & Lange, M. A. *J. Glaciol.* **34,** 128–135 (1988).
13. Huybrechts, P. *Ber. Polarforsch.* **99** (1992).
14. Lindstrom, D. *Paleoceanography* **5,** 207–227 (1990).
15. Lingle, C. S. & Clark, J. A. *J. geophys. Res.* **90,** 1100–1114 (1985).
16. MacAyeal, D. R. & Thomas, R. H. *J. Glaciol.* **32,** 72–86 (1986).
17. MacAyeal, D. R. *J. geophys. Res.* **97,** 595–603 (1992).
18. Engelhardt, H. *et al. Science* **248,** 57–59 (1990).
19. Kamb, B. *J. geophys. Res.* **96,** 16585–16595 (1991).
20. Alley, R. B. *et al. J. geophys. Res.* **92,** 8931–8940 (1987).
21. Boulton, G. S. & Hindmarsh, R. C. A. *J. geophys. Res.* **92,** 9059–9082 (1987).
22. Alley, R. B. *et al. J. geophys. Res.* **92,** 8921–8929 (1987).
23. Fortuin, J. P. F. & Oerlemans, J. *Ann. Glaciol.* **14,** 78–84 (1990).
24. Broecker, W. *et al. Clim. Dynam.* **6,** 265–273 (1992).
25. Thomas, R. H. *et al. Nature* **277,** 355–358 (1979).
26. Blankenship, D. D. *et al. Nature* **322,** 54–57 (1986).