Changes in the West Antarctic Ice Sheet

R. B. Alley and I. M. Whillans

The portion of the West Antarctic ice sheet that flows into the Ross Sea is thinning in some places and thickening in others. These changes are not caused by any current climatic change, but by the combination of a delayed response to the end of the last global glacial cycle and an internal instability. The near-future impact of the ice sheet on global sea level is largely due to processes internal to the movement of the ice sheet, and not so much to the threat of a possible greenhouse warming. Thus the near-term future of the ice sheet is already determined. However, too little of the ice sheet has been surveyed to predict its overall future behavior.

Changes in large ice sheets have affected global sea levels in the past, and may affect them in the future. The West Antarctic Ice Sheet (WAIS) has received special attention because it is different from the other ice sheets remaining on earth, and because, being grounded well below sea level, it may be prone to more rapid collapse. Indeed, paleontological analysis of subglacial sediments shows that the ice sheet was partially or completely absent at least once during the last 600,000 years (*1*).

The WAIS exhibits several styles of ice flow. Most of the ice volume is in the interior, and is called inland ice. It moves slowly, mainly by internal deformation. The flow of inland ice converges into fast-moving ice streams, which are separated by slow-flowing interstream ridges. The ice streams flow outward and eventually float as ice shelves. The various flow components are closely linked, such that a change in any one must propagate both up-flow and down-flow, throughout the ice sheet.

Early research focused on the ice shelves and on the marine nature of the ice sheet. The bed of the ice sheet is well below sea level and deepens toward its center (*2*); these features make the ice sheet prone to accelerating decay. Modeling and measurements show that there are longitudinal compressive forces in the ice shelves in addition to the tension normally expected for free flotation (*3*). This extra compression, or buttressing, acts up-glacier on the grounded ice streams and inland ice. Theoretical studies show that the magnitude of this buttressing is especially sensitive to changes in the melt rate under the ice shelves, and that, in turn, is sensitive to changes in oceanic circulation (*4*). In the most catastrophic scenario, an increase in ocean temperature would cause the ice streams to flow faster, thin, and float. The changes would begin in the ice shelves and propagate up-glacier, perhaps within a few centuries, such that the grounded ice would be thinned and sea level would be raised by up to 6 m.

R. B. Alley is at the Earth System Science Center and Department of Geosciences, The Pennsylvania State University, 306 Deike Building, University Park, PA 16802. I. M. Whillans is at the Byrd Polar Research Center and Department of Geological Sciences, The Ohio State University, 125 South Oval Mall, Columbus, OH 43210.

This simple picture was complicated by the discovery of fast ice streams within the grounded portion of the ice sheet (*5*). The ice streams are bounded by much more slowly moving ice, and are mainly heavily crevassed at the surface. However, for ice stream C, the crevasses are buried. This observation led to the suggestion that ice stream C had stopped recently, the crevasses being a relict from a former active phase of the ice stream (*6*). The stoppage of ice stream C while its neighbors continue to move rapidly points to an internal instability rather than some uniform change propagating up-glacier. Thus, there is not only the problem of why there are ice streams, but the possibility of major ongoing changes in the ice sheet. In this Article, we review studies that confirm these inferences and that have identified other rapid changes in the ice sheet and that demonstrate ice-flow instabilities.

Ongoing Changes

New work confirms the inference that ice stream C has stopped. On an ice-penetrating radar display, ice stream C looks like an ordinary, active ice stream, with abundant scattering from near-surface crevasses (*6, 7*). Satellite imagery shows streamlines similar to those of other ice streams, but more subdued (*8*). However, radar sounding and accumulation measurements show that the surface crevassing is buried by about 200 years of snowfall (*9*). Velocity measurements made with the use of Doppler satellites show the ice stream to be virtually stagnant (*10, 11*). Evidently ice stream C has stopped in the past few centuries and is now thickening at the surface accumulation rate, which is about 10 cm/year (*12*).

The neighboring ice stream B, in contrast, is thinning. Its discharge has been measured near its mouth, with the use of radar and repeat photogrammetry to measure the displacements of crevasses. The ice stream is draining ice 40% faster than replenishment by snow on its surface and in its catchment area. This amounts to an average thinning rate of 6 cm/year over the drainage basin (*12*), a large imbalance.

The imbalance in the ice stream B system is not distributed uniformly. Measurements of horizontal ice divergence indicate that thinning is concentrated at the head of ice stream B and locally reaches 1 m/year (*13*). Moreover, velocities measured with the use of Doppler satellites do not accord with flow directions inferred from mapped flow features (*11*). The end of the interstream ridge between the two major tributaries of ice stream B is especially anomalous. It would ordinarily be expected to be nearly stagnant, or perhaps be pulled along in the direction of the ice stream. However, it is moving sideways, probably due to a push from the vigorous ice stream to its south. Such irregular flow patterns may be the reason for observed contortions in surface-elevation contours (*14, 15*), difficult-to-understand crevasse patterns (*16, 17*), and contorted internal structures (*18*). There seems to be an interfingering of different flow regimes such that blocks or rafts of inland ice are being incorporated into the ice stream. The formation of rafts is believed to be associated with the headward migration of the ice stream at a rate of a few hundred meters per year (*19*). One such raft has been detected far down-glacier. The feature known as "ice rise a" (Fig. 1) has the visible and radar signature of slowly moving inland ice; however, it is moving quickly with the ice stream (*9, 20*). It seems that the flow of ice stream B has been and will continue to be time-variable.

Overall, ice stream B is thinning, but near the mouth of ice stream B the modern speed is much smaller (by about 100 m/year or 20%) than 10 years ago (*21*). This change indicates a thickening. The region of thickening continues down-flow to Crary Ice Rise (CIR: Fig. 1). This is a grounded ice dome within the Ross Ice Shelf. Today, the up-glacial end is thickening at about 0.8 m/year, while the other end is thinning at about 1.0 m/year (*22–24*). The net effect is that the ice rise is migrating up-flow in a complex way (*22*). The complexity is shown by a large raft of ice (40 by 7 km) that is being plucked from the side of the ice rise and carried off in the ice shelf (*25, 26*), and there are further changes in the ice shelf to the west of CIR (*27, 28*). The ice rise has been in its current location for 1000 years or less (*29*). These changes in the region of Crary Ice Rise may be a consequence of progressive grounding of the ice sheet, or they could be a contributing cause, because once the ice rise forms it becomes firmly fixed to the bed and blocks the outflow of the ice stream that runs into it (*24*).

Beyond the mouth of ice stream C, the recently stopped ice stream, the ice shelf is thinning. The stretching ice shelf is pulling on the ice stream, and has caused the grounding line to move 300 m up-glacier in 11 years (*30*).

The inland ice has been studied most carefully near Byrd Station, in the catchment of ice stream D. During the past 30,000 years, the ice flow has been roughly steady (*31*). But a small thickening began at the end of the last glaciation (*32*). This thickening has ceased, and at present ice outflow exceeds ice accumulation by 20%; as a result, the ice is thinning by about 3 cm/year (*33*).

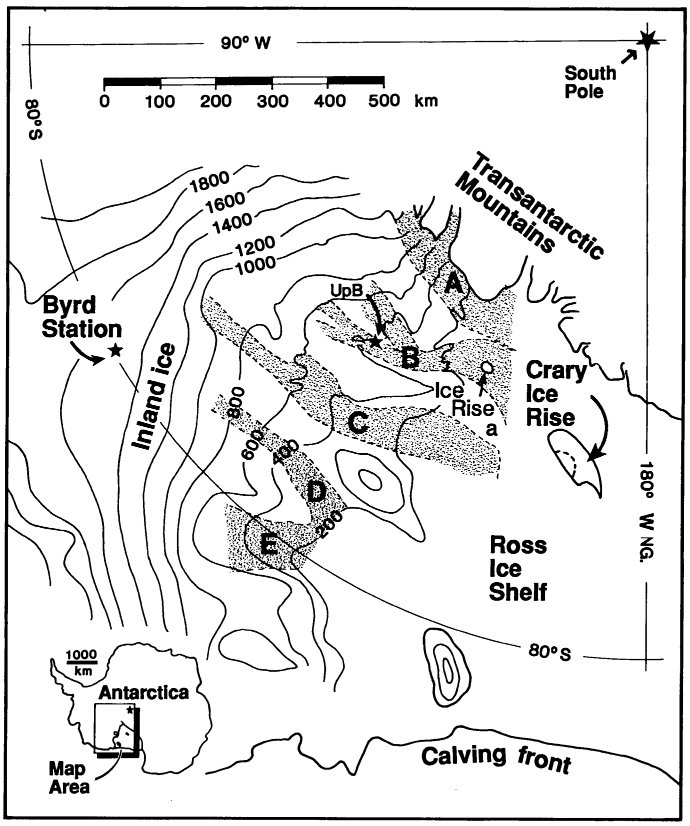

Fig. 1. The Siple Coast region of West Antarctica (*15, 63*); contour interval for elevations is 200 m.

Only certain parts of the ice sheet have been studied with precision (inland ice near Byrd Station, ice streams B and C, and their associated interstream ridges and the ice shelf). Thus, the net imbalance of West Antarctica is not well known. However, it seems appropriate to suspect that the remaining ice streams that flow into the Ross Ice Shelf also may be out of balance. There is a good possibility that the overall mass balance of West Antarctica is distinctly nonzero (*9, 13*).

Causes for Changes

The major external controls on the dimensions of an ice sheet are net snow accumulation, surface temperature, and marginal position (*34, 35*). A change in any of these controls must lead to changes in ice-sheet configuration. Models predict that for the WAIS, the time scale for such changes is thousands of years.

Data from ice-core records in Antarctica indicate that temperature and accumulation rate increased from about 15,000 to 10,000 years ago (at the end of the last global glaciation), and that they have been relatively constant since then (*36*). By itself, the increase in accumulation rate must have caused a thickening of the inland ice that should be largely completed by today. Warm ice flows more easily than cold ice, but a surface warming takes thousands of years to diffuse and be advected to the depths that control flow. Warming thus causes a delayed speeding and thinning of the ice, which should be occurring today (*37*). Thus, the climatic changes at the end of the last ice age should have caused a thickening and then an ongoing thinning.

From ~15,000 to ~5,000 years ago global sea level rose dramatically (*38*) owing at least in part to the collapse of the Northern Hemisphere ice sheets. Since then changes in sea level have been slow. The big rise caused much of the margin of the ice-age WAIS to float off its bed, which in turn caused thinning of the marginal regions and formation of the modern ice shelves (*39, 40*). This process must have also caused a wave of thinning to propagate inland over thousands of years; the inland ice is probably still responding (*34*).

In sum, the ordinary response to the environmental changes at the end of the last glacial period should be regionally coherent marginal thinning and retreat, together with inland thickening followed by thinning to the present. Glacial-geologic evidence from marginal areas shows that the ice sheet has thinned as much as 1000 m and the grounding-line has retreated as much as 1000 km coincident with the deglacial rise in global sea level (*40*). As noted above, results from Byrd Station and vicinity show the expected thickening followed by thinning. These measurements fit very nicely with the predicted response to the known external environmental forcing.

The changes in the ice streams do not fit this pattern. The changes may have been set on course by external events, but now, internal processes that are specific to ice streams must be playing a large role. Among the internal processes that may be significant are the controls on basal sliding and the transition from slow to fast flow at the ice-stream heads and margins.

Ice streams move much faster than inland ice despite having smaller basal shear stresses. This relation is illustrated in Fig. 2, which shows quantities along a flowline from the ice divide along ice stream B to the ice shelf. The driving stress (*41*), which causes ice flow, is usually about equal to the basal sheer stress. Paradoxically, it is a maximum where the ice is slow. Ice speed reaches a maximum in the ice stream, where the driving stress is small. Unlike the predictions of conventional theory (*42*), the speed of ice stream B is not linked to driving stress. Some other factor, such as basal lubrication, must be more important. The lubricant was first thought to be

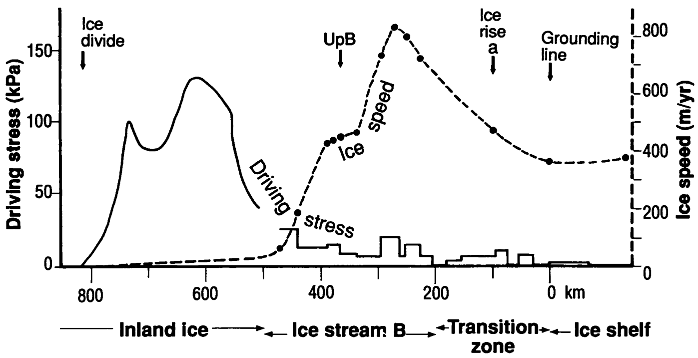

Fig. 2. Driving stress (solid), which is responsible for ice flow, and ice speed (dashed). Driving stress peaks where speed is small and vice versa, the peaks are about 400 km apart. The "transition zone" has been termed "ice plain," "ice apron," and "pseudo–ice shelf" (*63*).

water, which can separate the base of the glacier from its bed. The amount of lubrication is greater the more meltwater produced and the more difficult subglacial drainage. This view is supported by the presence of pressurized subglacial water under both the inland ice and ice stream B (*43–45*).

However, the lubricant may be more than just water. Porous, unconsolidated debris containing pressurized water has been detected under ice stream B with the use of seismic techniques and direct drilling (*44–46*). The pressure in the pore water supports most of the ice overburden. The debris is unconsolidated, and may be in motion (*47, 48*). The debris has been observed seismically to be nearly continuous over about 10 km by 10 km near the UpB camp on ice stream B (Fig. 1) and over a similar area about 200 km down-glacier (*49–51*). If the entire bed is covered by this debris, the frictional drag of the ice stream must be transmitted through it. This effect means that the ice is pulling the debris along, and moreover, that the velocity of the ice stream is controlled by the strength and thickness of this mobile debris (*47, 48*). If bedrock knobs project through the debris, then the frictional drag of the ice stream is divided between the low-strength debris and the knobs.

Debris in motion must have a source as well as a sink. At the source, the ice erodes its bed deeper. Geological and seismic-reflection studies show that the unconsolidated debris beneath ice stream B rests on and is derived from weak sediments of Tertiary age (*49, 51, 52*). Such a bed is easily erodible. The ultimate disposition of the debris is as submarine deltas, as has been reported for the mouth of ice stream B on the basis of seismic measurements (*49*).

If the bed is continuously mobile, the thickness and viscosity of the basal debris are crucial to the dynamics of the ice streams and the entire ice sheet. The behavior of the debris is sensitively tied to the mode of water transport and storage, as well as bed erosion, transport, and deposition. The viscosity of debris is extremely sensitive to the difference between ice overburden and water pressure, as well as to the amount of water or porosity. The water responsible for separating the glacier from its bed is produced by frictional dissipation and geothermal heat. It drains in irregular thin films over the debris and can be advected with the debris (*48, 53*). The velocity of the ice also increases with the thickness of the debris. In part, that depends on the balance between erosion of the bed and transport or deposition of till. Subtle changes in transmission of water or in debris erosion and transport would affect the thickness and viscosity of the debris, and have large effects on the dynamics of the overlying ice sheet (*54*). For example, the ill-defined boundary between the upglacier ends of ice streams B and C suggests that basal lubricant generated in the catchment of ice stream C has been diverted down ice stream B; if so, this would have contributed to the stoppage of ice stream C and the acceleration of ice stream B (*55*).

The transition from slow to fast flow at the lateral margins and at the upglacial origins of the ice streams is one of the most striking features of the WAIS and is a major puzzle. The transition is especially abrupt at the lateral margins of ice streams (Fig. 3), where the ice streams abut against the interstream ridges. Outside of the lateral margins, the ice flow is slower than about 10 m/year and can be accounted for entirely by internal deformation. The ice may be frozen to the substrate. This slow ice flows into the ice stream where it makes a sharp right-angle turn and increases speeds to more than 300 m/year (*56*). The ice is transformed from ridge-ice to ice-stream ice as it crosses through the boundary, which can be 5 km or less in width (*12, 20, 56, 57*).

The speed transition is not closely related to topographic features in the bed. Although the ice streams are roughly located over basal troughs, the margins in some places overlie basal deeps and in others, basal cross-slopes (*58*). In other ice sheets, the switch from inland ice to fast-flowing outlet glaciers occurs at a downstep in the

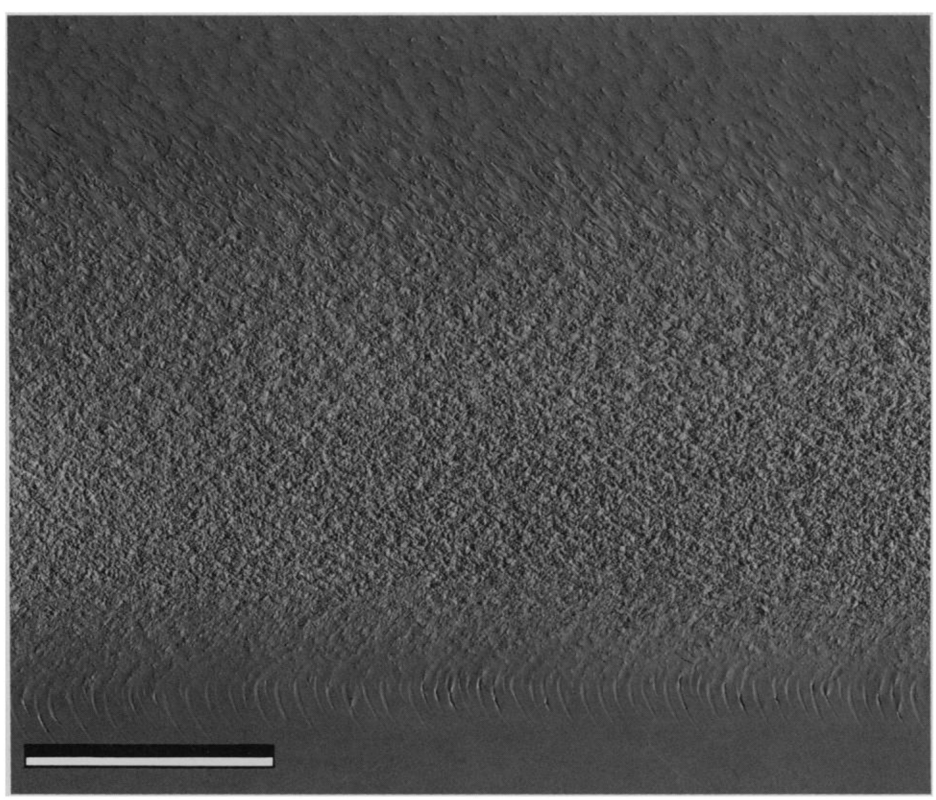

Fig. 3. Vertically taken aerial photograph of the northern shear margin of ice stream B. South is toward the top of the figure and solar illumination is from the right. The smooth region to the bottom is the slow-flowing ridge B/C. The zone at the top is ice stream B, flowing to the right at about 500 m per year. Most of the strike-slip shear occurs in the chaotic zone in the center of the figure. At the edge of ridge B/C is a series of *en echelon* hook-shaped crevasses. Photo 12-116. Scale is 1 km.

bed (*59*), but no such basal feature has yet been identified for the WAIS ice streams. The influence of bed topography on the nature of the ice streams appears to be weak.

The abrupt flow transition is most likely controlled by the onset of easy basal sliding, associated with the presence of abundant lubrication. This lubrication may be picked up by the ice as it attains some speed or drag threshold or may be injected under the incoming ridge ice from beneath the well-lubricated ice stream.

Another possibility for the occurrence of abrupt ice-stream margins is that the ice-stream ice may be softer than ridge ice, owing to different crystal orientation, temperature, or other factors. If this is the case, the softening effect must migrate into the ridge ice at about the same rate that the ridge ice is flowing into the ice stream.

The reason for the formation of rafts is not known. Perhaps large parts of the bed become lubricated all at once, or perhaps there are switches in ice stiffness or lubrication that form narrow shear margins and allow rafts to be pulled away from inland ice and ice rises. Along the ice stream, crevasse patterns suggest that the basal lubrication is not uniform (*17*), and these variations could be accentuated near the ice-stream head.

Conclusions

The WAIS is changing now, even before there could have been important anthropogenic effects. The slow thinning observed for the interior appears to be a continuing response to the major global changes at the end of the last glacial, but the complex changes occurring in the ice streams are not predicted by current models.

A widely held theory for the behavior of the WAIS is the ice-shelf buttressing theory (*60*). In this theory, the thin and fragile ice shelves hold back the grounded portions of the ice sheet through drag against islands and rock walls. Changes originating in the ice shelves propagate up-glacier. However, although the ice stream B system is mainly thinning, its mouth is thickening and slowing. This pattern is not accounted for by the classic buttressing model. Moreover, the near-zero longitudinal stress at DnB indicates that stresses from the ice shelf are unimportant in controlling flow styles at this site (*61*). The origin of many or most of the changes must lie within the grounded portion of the ice sheet, and especially within the ice streams.

Hypotheses to account for the observed ice-stream changes include changes in the generation and transport of water and basal debris, and variations in ice strength. These are mechanisms that are mainly internal to the ice sheet. Until these hypotheses can be tested and more fully developed it is not possible to predict the future course of the ice sheet.

The importance of the ongoing changes in West Antarctica to global sea level is not known. This is because such a small part of the ice sheet has been surveyed, and extrapolation of the current findings to the entire ice sheet would be very uncertain. However, the documented changes that are now occurring are very large, and so it seems probable that the ice sheet is contributing in some important way to the mass balance of the world's oceans, independent of any projected climatic changes.

REFERENCES AND NOTES

1. R. P. Scherer, *Global Planet. Change*, in press; also J. H. Mercer, *Nature* **271**, 321 (1978).
2. D. J. Drewry, Ed., *Antarctica: Glaciological and Geophysical Folio* (Scott Polar Research Institute, University of Cambridge, Cambridge, 1983).
3. K. C. Jezek, *J. Geophys. Res.* **89**, 1925 (1984); R. H. Thomas, *J. Glaciol.* **12**, 45 (1973).
4. C. J. van der Veen, *Ann. Geophys.* **4**, 45 (1986); C. S. Lingle, D. H. Schilling, J. L. Fastook, W. S. B. Paterson, T. J. Brown, *J. Geophys. Res.* **96**, 6849 (1991).
5. This discovery was made with the use of an ice-penetrating radar mounted in an aircraft; G. deQ. Robin, C. W. M. Swithinbank, B. M. E. Smith, *Int. Assoc. Sci. Hydrol. Assoc. Int. Hydrol. Sci.* **86**, 97 (1970).
6. K. E. Rose, *J. Glaciol.* **24**, 63 (1979).
7. S. Shabtaie, I. M. Whillans, C. R. Bentley, *J. Geophys. Res.* **92**, 8865 (1987).
8. R. A. Bindschadler and P. L. Vornberger, *Eos* **71**, 741 (1990).
9. S. Shabtaie and C. R. Bentley, *J. Geophys. Res.* **92**, 1311 (1987).
10. R. A. Bindschadler, D. R. MacAyeal, S. N. Stephenson, in *Dynamics of the West Antarctic Ice Sheet*, C. J. van der Veen and J. Oerlemans, Eds. (Reidel, Dordrecht, the Netherlands, 1987), pp. 161–180.
11. I. M. Whillans, J. Bolzan, S. Shabtaie, *J. Geophys. Res.* **92**, 8895 (1987).
12. I. M. Whillans and R. A. Bindschadler, *Ann. Glaciol.* **11**, 187 (1988).
13. S. Shabtaie, C. R. Bentley, R. A. Bindschadler, D. R. MacAyeal, *ibid.*, p. 137.
14. Elevations were obtained with the use of terrain-clearance radar measurements from an aircraft and ground control by Doppler-satellite tracking; see (*7, 15*).
15. S. Shabtaie and C. R. Bentley, *Ann. Glaciol.* **11**, 126 (1988).
16. As observed on aerial photography; see (*17*).
17. P. L. Vornberger and I. M. Whillans, *Ann. Glaciol.* **8**, 168 (1986).
18. There are complex internal radio-reflecting layers; D. L. Wright, S. M. Hodge, J. A. Bradley, T. P. Grover, R. W. Jacobel, *J. Glaciol.* **36**, 112 (1990).
19. In order to estimate the rate of migration of the ice stream, the surface elevation fluctuation, h, is approximated by part of a cyclic traveling wave

$$h = h_o\cos[2\pi(x - ct)/\lambda]$$

in which, x represents distance along the glacier, t is time, and c is the velocity of propagation of the fluctuation. Suppose that the changing elevation is due to the propagation of this wave. The rate of elevation change is then

$$\partial h/\partial t = h_o\,[2\pi c/\lambda]\sin[2\pi(x - ct)/\lambda]$$

which, by observation, reaches a maximum of 1 m per year. Other quantities are estimated as: wavelength, λ, 300 km; and amplitude, h_o, 100 m. Solving for c, one finds that the rate of headward migration is 500 m per year.
20. R. A. Bindschadler, S. N. Stephenson, D. R. MacAyeal, S. Shabtaie, *J. Geophys. Res.* **92**, 8885 (1987).
21. S. N. Stephenson and R. A. Bindschadler, *Nature* **334**, 695 (1988).
22. R. A. Bindschadler, E. P. Roberts, D. R. MacAyeal, *J. Glaciol.* **35**, 370 (1989).
23. The values are thinning or thickening plus or minus the rate of basal melting or freezing. D. R. MacAyeal *et al.* (*24*) suggested that basal melting and freezing are small and can be ignored.
24. D. R. MacAyeal, R. A. Bindschadler, S. Shabtaie, S. Stephenson, C. R. Bentley, *J. Glaciol.* **33**, 218 (1987); correction, **35**, 151 (1989).
25. This interpretation is based largely on an analysis of crevasse patterns (*26*).
26. R. A. Bindschadler *et al.*, *Ann. Glaciol.* **11**, 8 (1988).
27. There is a looped fold in flow features as observed by meteorological satellites (*29*).
28. G. Cassasa, K. C. Jezek, I. M. Whillans, *Ann. Glaciol.*, in press.
29. The time estimate is based on the analysis of temperature profiles near the center of the ice rise [R. A. Bindschadler, E. P. Roberts, A. Iken, *Ann. Glaciol.* **14**, 13 (1990)]. The steady-state basal temperature of a grounded ice rise like CIR is less than about −10°C. The temperature at the base of an ice shelf in phase equilibrium with sea water is about −2°C. The measured basal temperature under CIR of −4° to −5°C is between these values. The ice rise is thus believed to be a new feature, and its basal temperature is decreasing with time. The changes in CIR and the associated ice shelf have been in progress for at least 500 years, as evidenced by nonsteady features observed in the ice shelf [K. C. Jezek, *J. Geophys. Res.* **89**, 409 (1984); (*28*)].
30. R. H. Thomas *et al.*, *Ann. Glaciol.* **11**, 165 (1988).
31. Internal radar layers match isochrones calculated assuming constant accumulation rate and ice flow; therefore, both the flow and accumulation pattern have been unchanged [I. M. Whillans, *Nature* **264**, 152 (1976)].
32. There is more trapped air in deeper ice, indicating that this ice formed at lower elevations, where air density is greater, than the current ice surface; D. Raynaud and I. M. Whillans, *Ann. Glaciol.* **3**, 269 (1982).
33. Comparison of horizontal ice divergence with accumulation rate; I. M. Whillans, *J. Glaciol.* **18**, 359 (1977).
34. R. B. Alley and I. M. Whillans, *J. Geophys. Res.* **89**, 6487 (1984).
35. Other external controls are ordinarily expected to have even slower effects. These include erosion of the bed to expose a different bed type and changes in geothermal flux.
36. P. M. Grootes and M. Stuiver, *Quat. Res.* **26**, 49 (1986); C. Lorius, L. Merlivat, J. Jouzel, M. Pourchet, *Nature* **280**, 644 (1979).
37. I. M. Whillans, *J. Geophys. Res.* **86**, 4274 (1981).
38. R. G. Fairbanks, *Nature* **342**, 637 (1989).
39. This history is based on data from dated moraines and marginal lakes at the northwest corner of the Ross Ice Shelf (*40*) and on marine data [J. B. Anderson, in *West Antarctic Ice Sheet Initiative*, R. A. Bindschadler, Ed. (NASA Conf. Publ. 3115(2), National Aeronautics and Space Administration, Washington, DC, 1991), pp. 87–110.
40. G. H. Denton, J. G. Bockheim, S. C. Wilson, M. Stuiver, *Quat. Res.* **31**, 151 (1989).
41. Defined as ice density times the acceleration due to gravity times the ice thickness times the surface slope. For example, D. J. Drewry, *The Surface of the Antarctic Ice Sheet*, Sheet 5 of *Antarctica: Glaciologic and Geophysical Folio*, D. J. Drewry, Ed. (Scott Polar Research Institute, University of Cambridge, Cambridge, 1983).
42. For example, W. S. B. Paterson, *The Physics of Glaciers* (Pergamon, Oxford, 1981).
43. Drilling indicated that water is present at the bed at Byrd Station [H. T. Ueda and D. E. Garfield, *Int. Assoc. Sci. Hydrol.* **86**, 53 (1970)]; the water is at a pressure of 99 ± 1% of that calculated for the ice overburden [R. B. Alley, D. D. Blankenship, S. T. Rooney, C. R. Bentley, *Int. Assoc. Sci. Hydrol.* **170**, 81 (1987)]. Internal layers in the ice are folded as the inland ice flows over subglacial lakes [I. M. Whillans and S. J. Johnsen, *J. Glaciol.* **29**, 78 (1983)]. In ice stream B, seismic reflection work detected a low-strength bed (*44*). High-pressure subglacial water

was detected in direct drilling through ice stream B (*45*).
44. D. D. Blankenship *et al.*, *Nature* **322**, 54 (1986).
45. H. Engelhardt, N. Humphrey, B. Kamb, M. Fahnestock, *Science* **248**, 57 (1990).
46. D. D. Blankenship, C. R. Bentley, S. T. Rooney, R. B. Alley, *J. Geophys. Res.* **92**, 8903 (1987).
47. R. B. Alley, D. D. Blankenship, C. R. Bentley, S. T. Rooney, *Nature* **322**, 57 (1986).
48. ______, *J. Geophys. Res.* **92**, 8921 (1987).
49. R. B. Alley *et al.*, *Mar. Geol.* **85**, 101 (1989).
50. S. T. Rooney, D. D. Blankenship, R. B. Alley, C. R. Bentley, *J. Geophys. Res.* **92**, 8913 (1987).
51. ______, *Geological Evolution of Antarctica*, M. R. A. Thomson, J. A. Crame, J. W. Thomson, Eds. (Cambridge Univ. Press, New York, 1991).
52. R. P. Scherer, *Global Planet. Change*, in press.
53. R. B. Alley, *J. Glaciol.* **35**, 108 (1989); major flow in tunnels, or as groundwater seepage is discounted.
54. R. B. Alley, D. D. Blankenship, S. T. Rooney, C. R. Bentley, *J. Geophys. Res.* **92**, 8931 (1987); R. B. Alley, *Ann. Glaciol.* **14**, 1 (1990).
55. Basal fluids are mainly driven in the direction of the ice surface slope, with a smaller contribution from bed slope. At the boundary between the catchments of ice streams C and B the surface slope is now small, but the bed slopes down toward ice stream B. Perhaps a change in ice surface slope allowed the basal lubricant to be routed away from ice stream C and toward ice stream B. Such an event would enhance the speed of ice stream B and slow that of ice stream C, as observed [K. E. Rose, *J. Glaciol.* **24**, 63 (1979) (*16*).
56. P. L. Vornberger and I. M. Whillans, *J. Glaciol.* **36**, 3 (1990).
57. R. A. Bindschadler and T. A. Scambos, *Science* **252**, 242 (1991).
58. Radar depth sounding was by Shabtaie and Bentley (*15*).
59. N. F. McIntyre, *J. Glaciol.* **31**, 99 (1985).
60. R. A. Bindschadler, Ed., *West Antarctic Ice Sheet Initiative Science and Implementation Plan* (NASA Conf. Publ. 3115(1), National Aeronautics and Space Administration, Washington, DC, 1991), 53 pp.
61. Strain rate, $\dot{\varepsilon}_{ij}$, has been measured. It is linked to deviatoric stress, σ'_{ij}, through the flow law: $\sigma'_{ij} = B(\sigma)\dot{\varepsilon}_{ij}$, where $B(\sigma)$ is the viscosity. Full longitudinal stress is $\sigma_{xx} = 2\sigma'_{xx} + \sigma'_{yy}$. Combining: $\sigma_{xx} = B(2\dot{\varepsilon}_{xx} + \dot{\varepsilon}_{yy})$. From data in (*20*) (see figure 9), $\dot{\varepsilon}_{xx} = -1 \times 10^{3}\ \mathrm{yr}^{-1}$ and $\dot{\varepsilon}_{yy} = 2 \times 10^{-3}\ \mathrm{yr}^{-1}$ at the position of ice rise a, so $\sigma_{xx} \approx 0$.
62. D. J. Drewry, *The surface of the Antarctic Ice Sheet*, sheet 2 of *Antarctica: Glaciologic and Geophysical Folio*, D. J. Drewry, Ed. (Scott Polar Research Institute, University of Cambridge, Cambridge, 1983); R. A. Bindschadler *et al.*, *Ann. Glaciol.* **11**, 8 (1988).
63. Driving stress: smooth curve from D. J. Drewry, figure 5B in *Antarctica: Glaciologic and Geophysical Folio*, D. J. Drewry, Ed. (Scott Polar Research Institute, University of Cambridge, Cambridge, 1983); stepped portion left of 220 km from J. McDonald, thesis, Ohio State University (1990); to right of 220 km from (*15*). Speeds from J. McDonald; left of 220 km and right of 100 km from R. A. Bindschadler, S. N. Stephenson, E. P. Roberts, D. R. MacAyeal, D. R. Lindstrom, *NASA Tech. Memo. 100708* (1988).
64. We thank our colleagues in the Siple Coast Project. Support was provided by the U.S. National Science Foundation under grants DPP-8716016, DPP-8716447, and DPP-8915995. This is contribution 767 of the Byrd Polar Research Center.

Antigenic Diversity Thresholds and the Development of AIDS

MARTIN A. NOWAK,* ROY M. ANDERSON, ANGELA R. MCLEAN, TOM F. W. WOLFS, JAAP GOUDSMIT, ROBERT M. MAY

Longitudinal studies of patients infected with HIV-1 reveal a long and variable incubation period between infection and the development of AIDS. Data from a small number of infected patients show temporal changes in the number of genetically distinct strains of the virus throughout the incubation period, with a slow but steady rise in diversity during the progression to disease. A mathematical model of the dynamic interaction between viral diversity and the human immune system suggests the existence of an antigen diversity threshold, below which the immune system is able to regulate viral population growth but above which the virus population induces the collapse of the $CD4^{+}$ lymphocyte population. The model suggests that antigenic diversity is the cause, not a consequence, of immunodeficiency disease. The model is compared with available data, and is used to assess how the timing of the application of chemotherapy or immunotherapy influences the rate of progress to disease.

MUCH UNCERTAINTY STILL SURROUNDS THE PROCESSES governing the development of acquired immunodeficiency syndrome (AIDS), after an individual is infected with the human immunodeficiency viruses (HIV-1 and HIV-2). There is a long and highly variable incubation period for AIDS, with roughly 50 percent of male homosexuals developing the disease within 10 years after infection (*1*), and a slow but steady depletion of $CD4^{+}$ T-helper or inducer lymphocytes over this period in those who develop AIDS (*2*). The interaction between the viral population and the host's immune and other systems is very complex, with the virus being able to infect not only cells within the immune system but also a wide variety of other cell types in the brain, the gastrointestinal tract, the kidney, and other tissues (*3*).

Various explanations have been offered for the slow impairment of immune functions and the increased susceptibility of AIDS patients to opportunistic infections. These range from those based on the ability of the virus to kill $CD4^{+}$ cells, to those that invoke the presence of other infectious agents, such as mycoplasms, as necessary cofactors for the development of disease (*4*).

Understanding what is going on might seem to have been made more difficult by the discovery of great genetic diversity in viral isolates obtained either sequentially from the same infected individual or from different individuals (*5*). As a retrovirus, HIV lacks mechanisms that correct errors during replication, and the result is an error rate of about 10^{-4} per base, or one misincorporation per genome per replication cycle (*6*). Thus, each viral genome must be viewed as being different from any other, and viral isolates must be

M. A. Nowak, A. R. McLean, and R. M. May are in the Department of Zoology, University of Oxford, South Parks Road, Oxford OX1 3PS, UK; T. Wolfs and J. Goudsmit are in the Human Retrovirus Laboratory, Department of Virology L1-157, 1105 AZ Amsterdam; and R. M. Anderson (and R. M. May) are in the Department of Biology, Imperial College, London University, London SW7 2BB, UK.

*To whom correspondence should be addressed.