by Richard B. Alley

West Antarctic collapse—How likely?

Because most of the world's ice occurs in Antarctica, what might be the possible effects of global greenhouse warming on its ice sheets? In contrast to most mountain glaciers and Greenland, a warming trend is likely to cause an increase in snowfall on Antarctica, which would partially or totally balance the sea-level rise resulting from the melting of ice in the Northern Hemisphere. However, the possibility of dynamic instability in ice sheets exists, particularly for marine ice sheets. If this is the case, a marine ice sheet could collapse in one century or in a few centuries and cause sea level to rise rapidly.

Studies in West Antarctica show that some of its ice streams move very rapidly, apparently because of significant basal lubrication. The high ice-stream velocities result in local thinning of ice that might expand and bring about the collapse of the ice sheet. Another possibility is that a global greenhouse effect could warm the ocean waters circulating beneath the ice shelves and melt the ice at its base, even without melting the top surface of the ice. Again, this could lead to the catastrophic collapse of the West Antarctic marine ice sheet and possibly parts of the East Antarctic and Greenland ice sheets as well. (Ed.)

Introduction

Changes in global sea level over tens of years to hundreds of thousands of years are caused by changes in ice sheets. Only 20,000 years ago, during the Wisconsinan/Weichselian ice age, sea level was about 120 meters (m) lower than it is today. Deglaciation and sea-level rise occurred in rapid steps (Fairbanks, 1989) that suggest dynamic collapse of the ice sheets.

Today, the loss of the remaining ice sheets would raise sea levels about 75 m. The greenhouse warming expected over the next centuries is similar to the total warming from the last ice age to the present, and we must ask whether the remaining ice sheets can survive such a warming. The preliminary answer is that changes over the next centuries will be either small or quite large, depending on the dynamic behavior of the West Antarctic ice sheet.

Here I briefly review the present knowledge of the state of the ice sheets and how their surfaces will respond to greenhouse warming. I then summarize some results from the Siple Coast Project, a major effort to predict the dynamic behavior of the West Antarctic ice sheet.

Ice-sheet growth and decay

At present, mountain glaciers and small ice caps contain enough ice to raise sea level about 0.5 m if they melted, and the Greenland ice sheet would raise sea level about 5 m if it melted. Most of the world's ice occurs in Antarctica, where it is divided into two distinct ice sheets across the Transantarctic Mountains. The West Antarctic ice sheet would raise sea level about 5 m if it melted, and the East Antarctic ice sheet, about 65 m if it melted (NRC, 1985).

Snowfall on the ice sheets is equivalent to the evaporation of about 7 millimeters per year (mm/yr) of water from the world's oceans, but almost the same amount of water is returned to the oceans by the melting of ice. Small glaciers seem to be retreating today and contributing about 0.5 mm/yr to sea-level rise, and the large ice sheets seem to have little net effect on sea level. However, the uncertainties in our knowledge of ice-sheet mass balance probably are larger than 1 mm/yr of sea level (NRC, 1985).

Glaciers and ice sheets grow when snowfall exceeds melting, and they shrink when melting exceeds snowfall. In addition to snowfall and melting, ice flow is important because it transports the ice mass from regions where it builds up (accumulation) to regions where it melts away (ablation) or breaks off into marginal seas to melt (calving).

The balance between accumulation and ablation is a complicated meteorological problem that is poorly understood. A simple generalization gives us considerable insight into the problem, however. If the mean annual temperature is significantly above freezing, more snow will melt than falls. If temperatures are lowered, melting will decrease relative to snowfall and glacier growth will occur. But at very low temperatures (that is, $< -40^{\circ}\text{C}$), air cannot carry much water vapor, which causes both snowfall and glacier growth to be slow. Glacier growth thus should be fastest at some temperature near but below freezing, the growth decreasing to strongly negative values at higher temperatures and to weakly positive values at lower temperatures. This generalization ignores the importance of synoptic forcing, changes in cloudiness, cloud-physical processes, and other factors, which could cause different responses. However, this generalization does seem to work quantitatively in central East Antarctica and at least qualitatively in many places (Robin, 1977), so it is a good starting point until better relations become available.

At present, most mountain glaciers appear to be on the warm side of the optimum growth temperature. Warming following the Little Ice Age has caused them to shrink, and further warming should cause further shrinkage (Meier, 1984). Greenland appears to be near the optimum temperature, so a sufficiently large warming should cause it to shrink and melt in place over thousands of years. Antarctica, on the other hand, is mostly much colder than the optimum; temperatures in the interior average $< -50^{\circ}\text{C}$, and accumulation is < 30 mm/yr of water equivalent. Warming is likely to cause an increase in snowfall on Antarctica, which would partially or totally balance the sea-level rise resulting from the melting of ice in the Northern Hemisphere.

This present summary ignores the contribution of ice dynamics, but a simple examination of ice flow provides no cause for alarm. The East Antarctic and Greenland ice sheets are partially or totally frozen to their beds, and they move primarily by deformation within the ice mass. The rate of deformation increases exponentially with temperature, and ice velocity can increase further if the bed thaws and the ice

slides over the bed. However, most of the ice deformation and all of the sliding occur near or at the bed beneath ice that typically is 3,000 m thick. A warming trend takes thousands of years to penetrate to that depth and to begin to affect ice flow. Relative to temperature, the central East Antarctic ice sheet has just begun to respond to the end of the last ice age (Whillans, 1981) (fig. 1, curve c). Barring some dynamic instability of the ice sheet, then, we would expect Antarctic growth in a greenhouse-warmed world to balance partially or totally the melting of mountain glaciers and the Greenland ice sheet for many centuries at least.

Marine ice instability

The rapid increases in sea level during the most recent deglaciation (as fast as 24 mm/yr; Fairbanks, 1989) suggest the possibility of dynamic instability in ice flow, and this possibility is supported by glaciological theory. Theorists believe that the ice sheets most prone to instability are the marine ice sheets, that is, the ice sheets that have beds below sea (or lake) level and that float at their margins.

To understand this potential instability, consider figure 2. A fully developed marine ice sheet has three elements: an inland ice sheet, ice streams, and ice shelves. The inland ice or catchment looks like the central regions of East Antarctica or Greenland today. The ice is thick, exceeding 3,000 m commonly, and flows about 10 m/yr or less, primarily by means of internal deformation. Flow velocities and directions change only slowly with distance across the surface. The surface slope increases downglacier, which provides the increasing gravitational driving stress to transport the additional snow that accumulates along the flow.

Ice shelves form at the downglacier end of the ice sheet, especially in coastal embayments. They are extensions of the grounded ice sheet and are hundreds of meters thick, though they float on seawater in most places. Ice shelves spread under their own weight at a rate that increases with the cube of the ice thickness. Occasionally, ice shelves run aground on islands (fig. 2) and form ice rises or ice rumples. An ice shelf begins at the grounding line, where ice first floats free of its bed, and ends at a calving front, where icebergs break off. Between

Figure 1.—Ice-sheet response to climatic forcing, taken from Alley and Whillans (1984). Shown is the modeled response of the ice thickness at Dome C, East Antarctica, to different changes. Curve a, effect of a 10-percent increase in snowfall for an ice sheet that is free to advance at its margins; curve b, as in curve a but having the position of the margin fixed by deep ocean water so that the thickness increase causes the surface slope to steepen; curve c, effect of 7 °C warming; curve d, effect of marginal retreat such as might have been caused by melting of Northern Hemisphere ice sheets 15,000 yr ago; and curve e, sum of curves b, c, and d. Curve e is one possible scenario for ice-sheet evolution from the maximum of the last ice age (time 0) to the present (15,000 yr) and into the future in the absence of any further forcing after the deglaciation. The accumulation-rate increase at the end of the last ice age may have been larger than 10 percent and would, therefore, give a larger initial thickening than is modeled here.

Figure 2.—Cartoon of a typical marine ice sheet, such as West Antarctica (NRC, 1985).

Figure 3.—Antarctica. Shaded regions are below sea level and would remain below sea level following deglaciation and isostatic adjustment (Drewry, 1983).

these limits, the force required to move the ice shelf past the sides of the embayment and past ice rises and rumpled provides a back stress that has a stabilizing effect (see Paterson, 1981, p. 174–183).

Between the inland ice and the ice shelves lie the ice streams. These are regions of relatively thin ice (about 1,000 m thick) that have small surface slopes and driving stresses that decrease downglacier. Despite this, they move very rapidly (100–1,000 m/yr or faster) and accelerate downglacier. Ice streams flow between ridges of slow-moving ice that are similar to the catchment areas. Internal deformation of the ice in the ice streams is unable to account for their observed high velocities, so basal lubrication must also be involved. Basal resistance balances the driving stress for flow along much of the length of the ice stream, but near the grounding line, the back stress from the ice shelves becomes increasingly important (see MacAyeal, 1989).

Because the surface slope of ice streams and ice shelves is very gradual, a small change in ice thickness can cause a large change in the position of the grounding line where the ice begins to float. And if the bed deepens toward the center of the ice sheet, grounding-line migration may accelerate once it is initiated. (Isostatic sinking tends to cause the bed to deepen toward the center of a large ice sheet.) Suppose that some perturbation, such as a small rise in sea level or a small thinning of the ice, causes the grounding line to retreat toward the center of a marine ice sheet that has a bed sloping down beneath its center. Retreat will cause the new grounding-line position to have thicker ice than it had at the old position. Because the spreading rate of the floating ice (and of grounded ice near the grounding line) increases with the cube of the ice thickness, the spreading and thinning rate of the ice at the new grounding line will increase and cause more ice to float as the grounding line migrates up the glacier. In this manner, a marine ice sheet could collapse in one century or in a few centuries and could raise sea level rapidly (Weertman, 1974; Thomas and Bentley, 1978).

Broad regions of the former Laurentide and Fennoscandian ice sheets were marine, and it is likely that they collapsed during deglaciation (Hughes and others, 1985). Today, parts of the Greenland and East Antarctic ice sheets are marine. Our attention focuses on the West Antarctic ice sheet, however, because it is almost entirely marine in character (fig. 3).

West Antarctic research

Concern with the possible collapse of West Antarctica was raised by Mercer (1968), who noted that global sea levels during the previous interglacial period were 5 m higher than they are today, that the West Antarctic ice sheet contains ice equivalent to about 5 m of sea level, and thus that the ice sheet may have collapsed at that time. This hypothesis remains untested but has stimulated much research in the area. The efforts of the United States have concentrated on the region that feeds into the Ross Ice Shelf (fig. 3, 4). Other nations (especially the British, German, Norwegian, and Russian Antarctic programs) have focused on the ice feeding into the Filchner and Ronne Ice Shelves. Airborne-radar profiling by the U.S. National Science Foundation, the Scott Polar Research Institute (UK), and the Technical University of Denmark (Drewry, 1983) has provided surface and bedrock profiles for much of the ice sheet. Here I will concentrate on the studies in the Ross embayment because they are the most ambitious yet undertaken and because I am most familiar with them.

During the 1960s and 1970s, programs were begun to study the Ross embayment. A core drilled by the U.S. Army Cold Regions Research and Engineering Laboratory through the inland ice at Byrd Station, upglacier from ice stream D, provided a detailed climatic record back into the last ice age. When data from the core were combined with surface surveys, conducted primarily by researchers from The Ohio State University (USA), the core provided a much clearer understanding of the dynamics and recent history of the ice sheet than was available previously (Gow and others, 1968; Whillans, 1977).

Figure 4.—Ice streams crossing the Siple Coast into the Ross Ice Shelf (SCP, 1988). Ice stream C is shown dashed because it stopped moving rapidly about 200 yr ago.

Subsequent coring and geophysical and glaciological surveys on the Ross Ice Shelf by scientists from the Universities of Nebraska, Wisconsin, and Maine (USA), among others, led to the first general description of a large ice shelf (for example, Thomas and others, 1984). Glacial-geological evidence from the area (for example, Denton and others, 1989) indicates that during the most recent interglacial period the grounding line advanced substantially, possibly to the edge of the continental shelf. A rise in sea level, brought about by the melting of the ice sheets of the Northern Hemisphere, probably caused the retreat from this advanced position. Model results indicate that the retreat would have continued to the center of the ice sheet if the Ross Ice Shelf had not formed; however, back stress from the ice-shelf sides and from ice rises served to stabilize the grounding line near its present position (Thomas and Bentley, 1978).

These model results suggest a way in which projected greenhouse warming could cause the collapse of the West Antarctic ice sheet without initiating surface melting. The undersides of the ice shelves are in contact with ocean water and undergo slow, net basal melting. If warmed ocean waters were to circulate under the ice shelves and increase this melting, ice interactions with islands and with ice-shelf sides might be reduced sufficiently to initiate collapse.

The Siple Coast Project

With the catchment areas and the ice shelf reasonably well studied from a glaciological standpoint, it remained to address the middle ground: the ice streams and their interactions with the other two parts of the ice sheet. This is the task of the Siple Coast Project (SCP). The project was initiated in 1983 by scientists from the National Aeronautics and Space Administration's Goddard Space Flight Center (R. Bindshadler), The Ohio State University (I. Whillans), the University of Chicago (D. MacAyeal), and the University of Wisconsin (C. Bentley) (USA), supported by numerous students and colleagues. Since 1983, six more groups have joined the project, including my group.

The ultimate goal of the SCP is to predict the future behavior of the part of the West Antarctic ice sheet that flows in ice streams across the Siple Coast and adjacent regions into the Ross Ice Shelf. This goal cannot be achieved without the accomplishment of two other objectives: determining the configuration and mass balance of the ice sheet and determining the physical controls on ice flow (SCP, 1988).

Even simple mapping in the remote and inhospitable climate of Antarctica is difficult and time consuming, but the SCP has developed innovative techniques that make efficient use of the resources available. Light aircraft (Twin Otters) are used to place receivers for TRANSIT (or Doppler) satellite broadcasts. Up to 11 receivers track satellites simultaneously. Refinement of the observations then allows point locations to be determined to within a few meters in three dimensions (MacDonald and Whillans, 1988). Preliminary experiments with Global Positioning System (GPS) receivers are promising, and GPS should replace TRANSIT as further GPS satellites are launched.

Points on the ice sheet are marked by using photographic targets and radar reflectors. Aerial photogrammetry conducted over the targets allows mapping of such features as crevasses and ice-stream margins (fig. 5) with considerable accuracy. Crevasses reveal the stress state and locations where the basal lubrication changes, for example (Vornberger and Whillans, 1986). Identification of the same features on photographs taken a year or two apart allows calculation of the velocity field at many points (Brecher, 1986).

Increasingly, satellite imagery is coming to supplement or replace aerial photogrammetry. High-resolution imagery from the U.S. Land-

sat satellite covers ice streams D and E, and the French SPOT satellite can take off-nadir images reaching as far south as ice stream B (fig. 4). Such images reveal individual crevasses and can be used to reconstruct velocity fields (Lucchitta and Ferguson, 1986). Lower resolution images, such as those from the Advanced Very High Resolution Radiometer (AVHRR), average data over the individual crevasses but clearly reveal the patterns of surface roughness that distinguish ice streams from interstream ridges (Bindshadler and Vornberger, 1990).

Airborne use of ice-penetrating radar tied to the surveyed points allows measurement of surface altitude, bed altitude, and ice thickness along the flight paths, and it identifies grounding lines, crevasse zones including basal crevasses, regions where the ice is frozen to the bed or thawed at the bed, and other features (for example, Shabtaie and Bentley, 1987). Radars towed on sleds behind snowmobiles supplement this information from airborne sensors by concentrating on especially interesting areas, as well as by using different frequencies and pulse characteristics in order to obtain information that is not practical to obtain from the air (for example, Wright and others, 1990).

Additional, detailed ground-based measurements are made next in order to fill in the picture of the ice sheet that is emerging from this survey work. High-resolution seismic surveys provide data on ice anisotropy (Blankenship and Bentley, 1987), subglacial lubrication (Blankenship and others, 1986), and bedrock conditions (Rooney and others, in press). The ice anisotropy both records and affects the deformation of the ice. Bedrock conditions may channel flow and may contribute to basal lubrication, as discussed below. Passive-seismic monitoring of local ice quakes by using fiber-optic-linked geophones allows the location of regions of active surface crevassing, as well as the determination of how much of the energy of ice flow is expended in a brittle rather than ductile manner (Blankenship and others, 1987a). Resistivity, gravity, and magnetic surveys contribute to the picture of the ice and bed.

Boreholes can be melted through the ice by using hot water. The holes allow direct sampling of the bed and measurement of basal water pressure, ice temperature, and other important parameters (Bindshadler and others, 1990; Engelhardt and others, 1990).

Ice movement can vary at a variety of scales, and the variations reveal where the environment resists ice flow and how the ice responds to that resistance. Repeated measurements of detailed strain networks

Figure 5.—Vertical aerial photograph showing the transition from the ridge between ice streams B and C (smooth ice, lower right) to the chaotically broken ice of the margin of ice stream B (upper left). The curved marginal crevasses at the junction are typical. (Photograph courtesy of I.M. Whillans.)

followed by inversion of the data for basal resistances, may prove to be a valuable technique (Van der Veen and Whillans, 1989).

Rates of snow accumulation can be determined from shallow ice cores. Atmospheric testing of atomic bombs caused strong reference horizons of radioactivity in 1955 and 1963. By measuring the depth of these horizons and the density of overlying snow, the rate of snowfall can be determined (Whillans and Bindshadler, 1988). Analysis of deeper ice cores (fig. 6) for paleoclimatic data tells what the climatic forcing on the ice sheet has been in the past (Groote and others, 1990).

This list of techniques is far from exhaustive. An important aspect of the research is the development of a hierarchy of numerical models to test and expand the data, which will lead toward the ability to predict the future of the ice sheet (MacAyeal, 1989). The coordinated application of all of these techniques has led to more than 100 scientific contributions and is allowing the SCP to make rapid progress toward understanding the ice sheet.

Figure 6.—Ice coring by the Polar Ice Coring Office. Ice cores reveal past climatic forcing on the ice sheet.

Ice-sheet changes

Traditional wisdom would indicate that ice sheets change slowly over a period of centuries to milleniums. We have been startled to find that the Siple Coast region of West Antarctica is changing rapidly over a period of decades to centuries. Among the exciting results:

- Ice stream C virtually stopped moving about 200 yr ago, although it appears to have been a normal ice stream prior to that time. Now it is receiving more snowfall than it carries away and is building up. However, its grounding line is retreating upstream, probably because the spreading of the floating ice shelf downstream is not balanced by ice flow from upglacier, which causes thinning (Shabtaie and Bentley, 1987; Whillans and others, 1987; Thomas and others, 1988).
- Ice stream B appears to contain huge blocks of ice incorporated from the inland ice. Now it is transporting to the ice shelf 40 percent more ice than collects from snowfalls in its drainage area, and thus it is thinning. The thinning seems to be concentrated at the upglacier end where the incorporation of ice blocks occurs, and it locally exceeds 1 m/yr (Whillans and others, 1987; Whillans and Bindshadler, 1988; Shabtaie and others, 1988).
- The Crary Ice Rise complex, downglacier from ice stream B on the Ross Ice Shelf, exerts a strong stabilizing influence on the ice stream. The complex is thickening about 0.8 m/yr on its upglacier end and is thinning at a similar rate downglacier. It has been near its

present position as little as 1,000 yr or less. Ice velocities upglacier from Crary Ice Rise have decreased by 20 percent (100 m/yr) over the last decade (MacAyeal and others, 1987, 1989; Bindshadler and others, 1989, 1990; Stephenson and Bindshadler, 1988).

One possible explanation of these observations is that a change such as shrinkage of Crary Ice Rise caused ice stream B to accelerate and "capture" (Rose, 1979) the ice formerly draining into ice stream C. This caused ice stream C to stop and ice stream B to grow upglacier by incorporating into its head large blocks of inland ice. This excess ice caused Crary Ice Rise to grow and migrate upglacier, which increased the back stress on ice stream B and slowed it (Thomas and others, 1988). We now are testing this and other hypotheses. Note, however, that these dramatic imbalances occurred before greenhouse warming could have had much effect on the ice sheet. In view of the slow response of inland regions of large ice sheets to climatic forcing (fig. 1), it is possible that the observed changes represent a very complicated, ongoing response to the termination of the last global glaciation.

Ice-stream mechanics

Force-balance calculations for ice stream B show that most of the resistance to ice-stream flow arises at its bed, although the resistance probably is not distributed evenly on a scale of hundreds of meters to kilometers. In addition, the basal resistance becomes subsidiary to the back stress from the ice shelf and ice rises near the grounding line (Vornberger and Whillans, 1986; Alley, Blankenship, Bentley, and Rooney, 1987; MacAyeal, 1989). The resistance to flow is exceptionally low, however. The driving stress for ice-stream flow is similar to or smaller than that for flow of inland ice and interstream ridges, but ice streams move about 100 times faster than inland ice or interstream ridges. The bed of ice stream B (and probably of most or all ice streams) thus must be exceptionally well lubricated. However, prior to the SCP, the nature of that lubricant was not known, which hampered efforts to model the ice sheet.

A series of active-seismic experiments has identified what may be the lubricant of ice stream B. By using a high-frequency, high-resolution seismic system, Blankenship and others (1986; Blankenship, Bentley, and others, 1987) discovered that a region of ice stream B is underlain by a water-saturated, unconsolidated sediment layer a few meters thick that has very high water pressure and a high porosity. The layer is virtually continuous over a 10-km² area near a survey camp on the upglacier section of the ice stream, and it recurs over a similar area almost 200 km downglacier (Rooney and others, 1987, 1988; Blankenship and others, 1989).

The high water pressure must weaken this unconsolidated sediment, and such high porosity would be unlikely to be sustained in the subglacial environment unless active deformation is occurring. We thus hypothesize that the high ice-stream velocities arise from deformation within this subglacial sediment mass or till (Alley, Blankenship, Bentley, and Rooney, 1986, 1987; see fig. 7). Such deformation has been observed beneath smaller glaciers, but this suggestion is a radical departure from the usual rigid-bed sliding models previously applied to Antarctic ice streams.

Rapid subglacial till deformation must lead to rapid sediment flux to the grounding line and must cause deposition of a soft-sediment delta or debris apron there. We estimate that such a deposit would have grown to some tens of kilometers in length since the last glacial period and would support only a gentle slope on the upper surface of the ice above these sediments (fig. 8). The discoveries by Shabtaie and Bentley (1987) that the grounding line of ice stream B is nearly 100 km farther downglacier than suspected previously and that the intervening "ice plains" have a very low ice-air surface slope and are underlain in places by unconsolidated sediment lend credence to our ideas. Indeed, subsequent seismic surveys have shown internal structures in the subglacial sediments of the ice plains that are consistent with our model

Figure 7.—Cartoon of models of the bases of ice streams. The traditional view is shown in model c, where the base of the ice slides over a rigid bed. New evidence favors model a, deformation within an unconsolidated till, or model b, till deformation plus sliding.

(Blankenship and others, 1989). Furthermore, we have been able to simulate the ice-stream profile by using a numerical model based on this deforming-bed hypothesis, something that was not possible previously by using any physically based, rigid-bed models (Alley, Blankenship, Rooney, and Bentley, 1987, 1989).

A drilling program conducted by B. Kamb and coworkers from the California Institute of Technology (USA) has verified the seismically determined water pressure beneath ice stream B. They found that the bed is a diamicton similar to a glacial till and is soft enough to be penetrated readily by a rigid probe (Engelhardt and others, 1990). Direct verification of subglacial deformation is not yet available, but most evidence so far points to this as the mechanism of ice-stream flow.

If verified, the deforming-bed hypothesis suggests that the resulting rapid sedimentation at the grounding line may help prevent grounding-line retreat and ice-sheet collapse. However, the ongoing erosion that supplies this sediment is deepening the bed at a few tenths of a millimeter per year (Alley, Blankenship, Bentley, and Rooney, 1987), which may destabilize the ice sheet on long time scales of 10^5 – 10^6 yr.

Further work

The exciting discoveries about the current rapid changes in the West Antarctic ice sheet lend urgency to our investigations. We really do not understand why these changes are occurring, and we cannot tell with confidence whether they will become widespread. However, if the regions undergoing local thinning of ice on a scale of nearly 1 m/yr were to expand to include large areas of the ice sheet, catastrophic collapse of the ice sheet would follow rapidly.

Our increasing understanding of ice-flow mechanics in the region may allow us to predict how ice-sheet collapse would proceed once it was initiated, but the mechanism of initiation—which already may have begun or which may be caused by human activities—remains enigmatic. This lack of understanding of the possible triggers for ice-sheet collapse stems from the interdisciplinary nature of the problem. The interaction of ice shelves with the ocean, which is itself affected by sea ice, atmospheric processes, and geologic processes, is likely to control any initiation.

We thus are in the planning stages for a new effort that will follow the SCP but will give proper attention to the systems around the ice

sheet, as well as to the ice sheet itself. This effort will be called SeaRISE (Sea-level Response to Ice Sheet Evolution; see Bindshadler, 1990). It will attempt to reach the SCP goal of predicting future ice-sheet and sea-level changes, not only for the West Antarctic ice sheet but also for the marine sections of the Greenland and East Antarctic ice sheets.

In summary, any dramatic changes in sea level on time scales of immediate human interest will come from ice sheets. Greenhouse warming probably will cause melting of mountain glaciers and may cause melting of the Greenland ice sheet, but it probably will increase snowfall on the Antarctic ice sheet, which would partially offset any sea-level rise. However, warmed ocean waters may circulate beneath the ice shelves and cause dynamic collapse of the West Antarctic ice sheet and possibly parts of the East Antarctic and the Greenland ice sheets as well. The Siple Coast Project finds that parts of the West Antarctic ice sheet are changing rapidly now. We also are gaining the ability to predict how the ice sheet will respond to marginal forcing, such as an acceleration of ice-shelf flow caused by enhanced basal melting. An interdisciplinary effort will be needed to predict such enhanced basal melting and the course of ongoing changes that began sometime in the past. This effort is in the planning stage, but until such an effort is complete, we do not know how likely it is that changes in ice sheets will cause rapid sea-level rise.

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Figure 8.—Cartoon of an ice stream on a deforming bed as the ice enters the sea. Subglacial sediments will be deposited at the grounding line to form a soft-sediment deposit that will support only a gentle slope on the upper surface of the ice above these sediments. Foreset-bed angles may be about 1° but are shown here with great vertical exaggeration. In addition, no bedrock relief is required at the grounding line, as shown. Geophysical evidence suggests the existence of deposits similar to this beneath ice stream B.

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