Decoding the dipstick: Thickness of Siple Dome, West Antarctica, at the Last Glacial Maximum

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ABSTRACT

Ice thickness in West Antarctica at the Last Glacial Maximum (LGM) is poorly known, yet is key information for understanding ice streams and interpreting ice cores. Although trim lines, moraine limits, and exposure-age dating provide geologic constraints on ice thickness near the Transantarctic Mountains and in Marie Byrd Land, lack of exposed bedrock hampers traditional geologic methods in a central, \(-2 \times 10^6 \) km\(^2\) region. Here we infer ice-sheet thickness changes in the central Ross Sea Embayment by using a transient ice-flow model to find combinations of accumulation-rate and ice-sheet thickness histories that match the depth-age relationship and the measured layer-thickness pattern in the Siple Dome ice core. After we reject unreasonable accumulation-rate histories, the remaining history pairs indicate thinning of 200–400 m since the LGM. Our estimate is lower than previous reconstructions that were constrained by geologic evidence from the Transantarctic Mountains and by marine data from the Ross Sea floor, which indicate that a grounded ice sheet extended to the continental shelf margin during the last glacial period. Low surface elevations in the central Ross Sea Embayment during the LGM do not preclude thicker ice along its boundaries. However, if this grounded ice sheet came over 1000 km from interior West Antarctica, as is usually assumed, then it had very low surface slope, requiring a very slippery bed. Alternatively, the grounded Ross Sea ice might have flowed from the Transantarctic Mountains and western Marie Byrd Land.

Keywords: Antarctic Ice Sheet, ice cores, ice streams, glaciology, sea level, climate.

INTRODUCTION

The grounding line of the West Antarctic Ice Sheet has retreated significantly since the Last Glacial Maximum (LGM) (Conway et al., 1999), but debate continues on the past volume and configuration of the ice sheet and its contribution to lower global sea level during the LGM (Bentley, 1999). Nunataks are absent in the central Ross Sea Embayment (Fig. 1), precluding traditional geologic “dipstick” measurements of past ice thickness from trim lines or exposure-age dating. Instead, we use an ice-flow model to find combinations of past accumulation rate and ice thickness that replicate the measured depth-age scale for the ice core at Siple Dome (Fig. 1). By selecting the most reasonable accumulation-rate history based on evidence from Siple Dome and other Antarctic ice cores, we also select the corresponding ice-thickness history.

Downward ice-flow velocity decreases with depth. This strain-rate pattern always causes vertical thinning and horizontal stretching, so an annual layer decreases in thickness over time. In an ice sheet that has thinned, an annual layer at a given depth is older than a layer at the same depth in an ice sheet that has not thinned (under the assumption that both ice sheets have the same accumulation rate and current ice thickness). The layer in the thinning ice sheet took longer to reach its present location because it had to travel farther downward from its point of origin on the surface, which was at a higher elevation when the snow was deposited; it also underwent more strain thinning in that time. However, if accumulation rates were higher in the past, then greater initial thickness of those layers could counter the effect of ice-sheet thinning. Ice-sheet thinning and/or lower accumulation rate in the past would produce relatively thin annual layers today. In addition, vertical velocity at mid-depths is lower under an ice divide than on the flanks of an ice sheet (Raymond, 1983), so past ice-divide positions also affect the depth-age profile observed today. Multiple combinations of past thickness \(H(t)\), accumulation rate \(b(t)\), and ice-divide position can produce the same depth profiles of age and annual-layer thickness. Short-term variations in layer thickness arise primarily from variations in accumulation rate, which affect layer thickness only at the time of deposition, whereas ice-sheet thickness changes and divide-position changes have a smoothly varying effect on layers through the entire depth of the concurrent ice sheet. Rapid os-
cillations in accumulation rate are preserved in the layer-thickness record, whereas effects of oscillatory changes in ice-sheet thickness that occur over times shorter than a characteristic time scale $H/b$ tend to average out.

The accumulation rate $b(A)$ when a layer of age $A$ was deposited, and the thickness $h(A)$ of that annual layer today, are related by the depth-age scale $z(A)$ and the thinning function $\Lambda(A)$ which incorporates total vertical strain, i.e., $b(A) = \Lambda(z(A))/h(A).$ We calculate $\Lambda(A)$ with a one-dimensional transient flow model in which vertical strain rate is uniform with depth in the upper part of the ice sheet. Below height $h$ (measured from the bed), strain rate decreases linearly to zero at the bed (Dansgaard and Johnsen, 1969). Comparison of the Dansgaard-Johnsen flow model with a steady finite-element flow model of Siple Dome (Nereson et al., 1998) indicates that using $h = 0.7H$ at the divide and $h = 0.25H$ on the flanks yields a good approximation of the present-day flow field. This simplified model allows investigation of transients in ice-sheet thickness, accumulation rate, and ice-divide position.

APPLICATION TO SIPLE DOME

During 1997–1999, an ice core to bedrock was recovered from Siple Dome. The site is 0.5 km south of the present divide location, where thickness is $H = 988$ m (ice equivalent). Taylor et al. (2004) established a depth-age relationship (Fig. 2A) by counting annual layers in the upper 514 m and by correlating measurements of occluded CH$_4$ with measurements from the well-dated GISP2 ice core from Greenland. The layer-thickness profile (Fig. 2B) is the gradient of depth vs. age. Because age errors should be correlated among methane-age–determined control points, large spurious temporal gradients in layer thickness are unlikely; we expect that the thick layers between 9 and 12 ka—inferred from the change in gradient of the depth-age relationship during that interval—are real. Annual-layer thickness at the surface (0.12 m ice equivalent) is consistent with the accumulation rate over the past 1 k.y. (Kreutz et al., 1997), and the characteristic time scale $H/b$ is $\sim 8$ k.y.

We represent divide location and migration through the Dansgaard-Johnsen flow-model parameter $h(r).$ Conditions of pure divide flow exist only at locations that are within a fraction of the ice thickness from a divide; vertical velocity at the core site 0.5 km (0.5$H$) south of the present divide is now transitional between divide flow and flank flow. Radar-detected internal layers (assumed to be isochrones) are arched upward beneath the divide because of reduced downward velocity there (Raymond, 1983); asymmetry of these arches indicates that the divide has migrated northward about 0.3 km in the past 1 k.y. (Nereson et al., 1998). Divide location histories that are consistent with this migration encompass divide flow ($h = 0.7H)$ at the core site at 1.5 ka, and transitional flow ($h = 0.5H$) from 1.0 ka to the present (Fig. 2C). Because conditions are less certain for earlier times, we investigated two possibilities prior to 4 ka: (1) divide far from the core site (flank flow, $h = 0.25H)$, solid blue curve), and (2) divide always near the core site (transitional flow, $h = 0.5H,$ dashed red curve). When we assumed that neither thickness nor accumulation rate changed since 25 ka, both divide histories produced poor matches between model and measurements; in Figure 2A (where line colors and patterns match divide migration histories in Fig. 2C), modeled ages were consistently too young. Except for a short interval near 10 ka, both models produced layers that are too thick (Fig. 2B). It is unlikely that rapid fluctuations in thickness or divide motion leave a record in the depth-age profile, so we specified relatively simple thickness and divide-migration histories. Then we adjusted the accumulation-rate histories until the model reproduced the measured age and layer-thickness profiles, as shown by thick gray lines in Figures 2A and 2B. For example, if the ice thickness has not changed, and if we accept either of the ice-divide histories in Figure 2C, then we must also accept the corresponding accumulation-rate history in Figure 2D.

LGM reconstructions of West Antarctica have ranged from little or no thinning at Siple Dome (Drewry, 1979) up to 1200 m of thinning (Stuiver et al., 1981). Analysis of radar-detected internal layers and surface-strain measurements (Nereson and Raymond, 2001) showed that Siple Dome has not changed appreciably in thickness since 2 ka. We next investigated scenarios in which Siple Dome was thicker during the LGM, and thinned to its present-day elevation over the period 16–2 ka (Fig. 3A). In all cases, the divide was distant from the core site prior to 4 ka (solid blue curve in Fig. 2C). Thickness-accumulation pairs that yield acceptable matches between modeled and measured age and layer-thickness profiles require progressively higher accumulation rate during the LGM and early Holocene to offset the effects of progressively more ice-sheet thinning (Fig. 3B). We also explored sensitivity to thinning-onset time. Figure 3C shows that thinning starting at 16 ka or at 8 ka; corresponding accumulation-rate histories (Fig. 3D) were not significantly different for just 200 m of thinning, but for larger thickness changes, delayed thinning onset can require as much as 50% more accumulation.
DISCUSSION

Additional information is necessary to distinguish between the effects of climate histories and thickness histories. Evidence for thicker ice in the Ross Sea Embayment during the LGM comes from dates of elevated moraines alongside outlet glaciers through the Transantarctic Mountains (Denton et al., 1989), from nunataks now being exposed in the Ford Ranges (Stone et al., 2003), and from geophysical measurements and models of radar-detected layers in Roosevelt Island (Conway et al., 1999). In addition, sediment cores from the western Ross Sea (Anderson et al., 2002) indicate that grounded ice extended out to the continental shelf (≈1000 km beyond its present location) during the last glacial period. Assuming that this grounded ice came from the interior of West Antarctica and that slopes of $\approx 10^{-3}$ (typical for present-day ice streams) were necessary for the ice sheet to extend that far, elevations near the present grounding line would have been $>1000$ m above the LGM sea level, or $850$ m thicker than at present (similar to dashed magenta curve in Fig. 3A).

In contrast, ice-core records suggest more modest thinning. The elevation of Taylor Dome (Fig. 1) has not changed significantly since the LGM, so temperature variations recorded by stable isotopes in the Taylor Dome ice core were caused primarily by changes in climate (Steig et al., 2000). Assuming that climatic temperature trends were similar at Taylor Dome and at Byrd (Fig. 1), Steig et al. (2001) attributed differences in the two stable isotope trends to elevation changes at Byrd, finding $\approx 200$ m of thinning there since 8 ka. However, in simple ice sheets, changes in thickness far downstream and near the margins are generally much larger than changes near an ice divide. Since Byrd Station is much closer to the primary West Antarctic ice divide, Steig et al. (2001) could not rule out much larger thickness changes near Siple Dome in the center of the Ross Sea Embayment. We also compared stable isotope trends in the Taylor Dome and Siple Dome cores. The elevation history we derive in this way for Siple Dome (see dotted red curve, Fig. 3C) shows little or no change between 12 and 8 ka, followed by thinning of $100-200$ m, to reach the present-day elevation by 2 ka.

We put additional constraints on the thickness history by excluding improbable accumulation-rate histories:

1. We expect the accumulation rates at Byrd and Siple Dome to show broadly similar trends through the deglaciation; as mid-tropospheric sites, both were subject to similar storm tracks. Thickness changes of $\approx 200$ m at Byrd (Steig et al., 2001), where the ice is $>2$ km thick, are too small to seriously affect the accumulation rates inferred from annual-layer thicknesses in the Byrd ice core. The accumulation rates inferred by Hammer et al. (1994) were less than half present-day values between 50 and 18 ka and rose uniformly toward modern values between 18 and 12 ka. Since 12 ka, the accumulation rate has been relatively constant. Hence we have low confidence in the no-thinning history for Siple Dome (Fig. 2D), because, apart from the high excursion near 12 ka, the accumulation rate continually rises from the LGM to present in this history. We also have low confidence in histories with more than $400$ m of thinning, because the high accumulation rate from 17 to 14 ka (Figs. 3B, 3D) is not evident in the Byrd record.

2. The accumulation rate was at most 20% higher (relative to today) between 12 and 9 ka at both Taylor Dome and Byrd. Therefore, we have low confidence in histories that require a very high accumulation rate at Siple Dome at that time, i.e., thickness changes of $>400$ m (or $>200$ m if the thinning occurred since 8 ka; Figs. 3C, 3D).

3. The accumulation rate probably followed similar trends at Taylor Dome and Siple Dome after the Ross Sea opened ca. 7 ka (Conway et al., 1999). After doubling between 8 ka and 6 ka, the accumulation rate at Taylor Dome then increased slowly by a further 20% to the present (Minnin et al., 2004). At Siple Dome, this Holocene rise is matched best with thinning of $400$ m or less (Figs. 3A, 3B). However, if all thinning occurred since 8 ka (Figs. 3C, 3D), even $400$ m of thinning is unlikely.

We can also constrain the divide history. Even in the absence of thinning, a divide that was always nearby (dashed red curve, Fig.
2C) requires an accumulation rate (dashed red curve, Fig. 2D) as much as 30% greater than a divide that was distant from the core site prior to 4 ka (solid blue curves). Additional model experiments showed that divide proximity has an even greater impact when thinning also occurs. For example, with 400 m of thinning since 16 ka, the accumulation rate had to exceed the modern value from 17 ka to 9 ka, and at 11 ka it had to be almost double the modern value. Even 200 m of thinning requires an implausibly high accumulation rate (30% above modern value) at 16 ka, and an improbable peak (60% above modern value) at 11 ka. These results suggest that the divide was often farther than 1 km from the core site prior to 4 ka.

Knowing ice thickness at the LGM is key to understanding the long-term behavior of ice streams; if Siple Dome was only 200 m thicker, then ice streams may have been active near Siple Dome at the LGM. Correct interpretation of the stable isotope proxy temperature record in ice cores from the West Antarctic Ice Sheet in terms of climate change also requires knowledge of the ice-sheet surface height. Our new method has converted the Siple Dome depth-age record into the first dipstick for LGM ice in the central Ross Sea Embayment. Our preferred solutions show 200±400 m of thinning, as indicated by the gray shaded areas in Figure 3. An independent accumulation-rate history at Siple Dome would allow even tighter constraints on the thickness history. Using a model for firm compaction and grain growth, a temperature history inferred from δ18O in the Siple Dome core, and measured bubble number densities in the core, M.K. Spencer (2004, personal commun.) suggests that accumulation rate increased slowly from 6 ka until present, if confirmed, this result would support our findings of modest thinning.

Our estimate of 200–400 m is higher than the change of only ~100 m inferred from thermo-mechanical simulations of ice streams by Parizek and Alley (2004). Reconstructions constrained by geologic data from the Transantarctic Mountains have been unable to tightly constrain past ice thickness in the center of the embayment. Some reconstructions with slippery beds and ice streams, or even floating ice (e.g., Drewey, 1979; Denton et al., 1989, minimum reconstruction), showed less thinning than our result, while other more widely accepted reconstructions, with ice streams of restricted length, showed much more; e.g., Denton and Hughes (2000) suggested 700 m thicker ice at Siple Dome. Furthermore, if the constraining geological data are incorrect, then the preferred reconstructions must also be reassessed. By revising an extrapolation of a key moraine, Anderson et al. (2004) suggested that the ice sheet near Hatherton Glacier (Fig. 1) was ~300 m lower during the LGM than previously thought. Incorporating this revised elevation in the Denton and Hughes (2000) reconstruction of the Ross Sea Embayment would produce a Siple Dome that was only 500 m thicker during the LGM. Thus, the geologic and glaciologic evidence may be converging.

Low surface elevations in the central Ross Sea Embayment during the LGM do not preclude thicker ice along the western and eastern boundaries (e.g., see minimum reconstruction of Denton et al., 1989), but an ice sheet of such low surface slope could extend 1000 km out to the continental shelf only if it was grounded for insufficient time to thicken, or if its bed was very slippery. A slippery bed could preclude thick ice near Siple Dome over longer time scales. Alternatively, the grounded ice in the Ross Sea might have come from the Transantarctic Mountains and western Marie Byrd Land, rather than from the interior of West Antarctica as is usually assumed.

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