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THE FLOW REGIME OF ICE STREAM C AND HYPOTHESES CONCERNING ITS RECENT STAGNATION

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The downglacier part of ice stream C, West Antarctica largely stagnated over the last few centuries, while upglacier regions continue to flow vigorously. We summarize hypotheses for explaining this behavior and suggest that the stagnation is due to a combination of water-diversion, thermal processes at the bed, and sticky spots. Stagnation likely occurred near Siple Dome before the entirety of the downglacier part slowed. Numerous data sets show that the slow-moving part of the ice stream is restrained largely by small, localized basal "sticky spots". The sticky spots are separated by extensive regions of soft till containing high-pressure liquid water. Near the transition from fast-moving well-lubricated ice to slow-moving ice, a hydrologic potential map indicates that basal water flowing in from the catchment is diverted away from the slow-moving ice to ice stream B. This diversion could have been caused by a flattening of the surface slope over time in response to the headward growth of ice stream C drawing down the inland ice. Previous mass-balance estimates indicate that the combined B-C drainage most likely is thinning slowly, similar to the rest of the Siple Coast.

1. INTRODUCTION

The dynamics of the West Antarctic Ice Sheet (WAIS) is dominated by fast-flowing ice streams that evacuate inland ice rapidly through three main ice-drainage systems to the ocean. The six ice streams that flow westward into one of these drainage systems, the Ross Sea Embayment (Plate 1), are separated by intervening ridges of slow moving ice. High shear strain rates at the lateral margins of the active ice streams result in long, linear, crevassed zones that are clearly visible from the air, in satellite images [e.g. *Bindschadler and Vornberger*, 1990], and in radio echo-sounding profiles [e.g. *Bentley*, 1987].

The first maps of the Ross ice streams came from airborne surveys and radar by *Robin et al.* [1970], who identified ice stream B and located several "pseudo ice shelves" either along ice stream C [interpre-

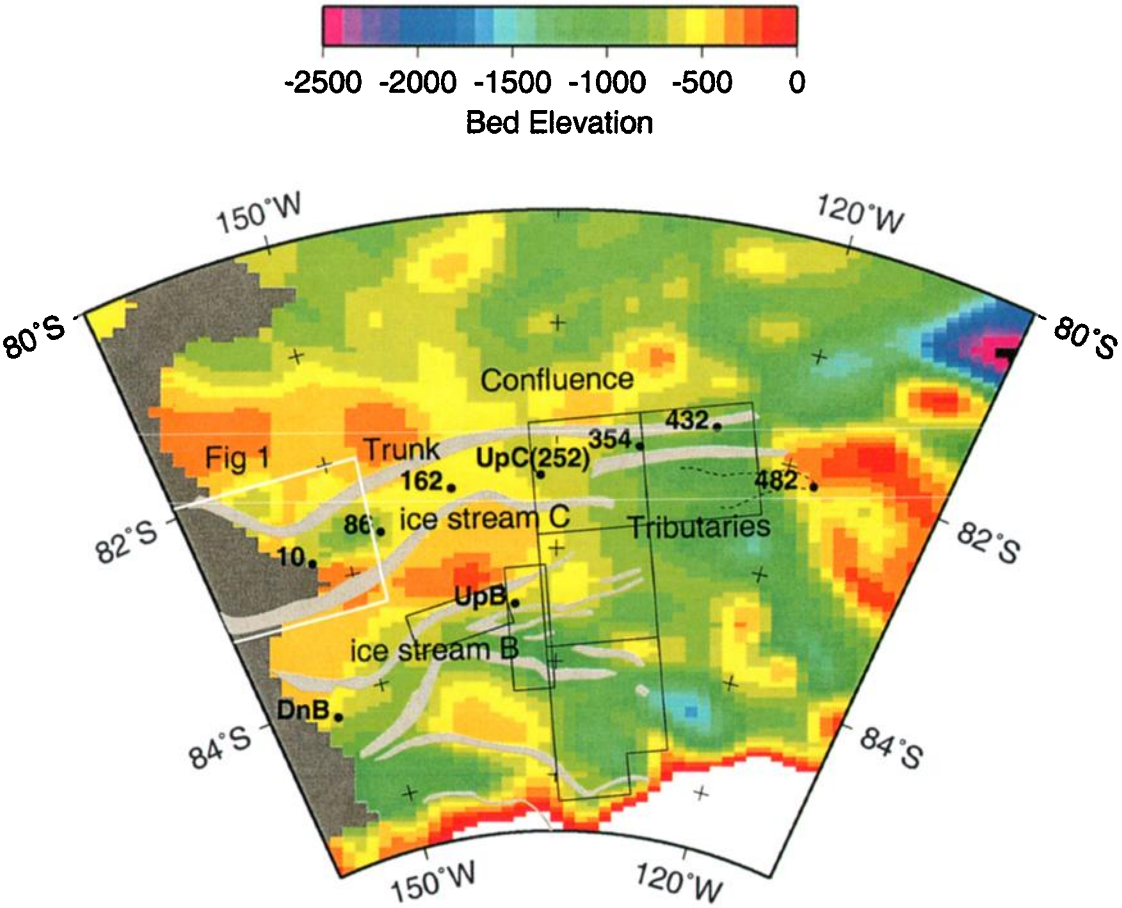

Plate 1. Map of the Siple coast region of West Antarctica showing the outlines of ice stream A, B, and C (from *Shabtaie et al.* [1987]) and the outline of the southern tributary of ice stream C (from *Hodge and Doppelhammer* [1996]; *Anandakrishnan et al.* [1998]). The bed elevation is based on radio-echo sounding data flown at a grid-spacing of 50 km (courtesy of J.L. Bamber). Flow is from the east to the Ross ice shelf on the west. The numbered sites on ice stream C are locations of microearthquake monitoring, where the site-number is the approximate distance in kilometers from the grounding line (which is from *Shabtaie et al.* [1987]). Upstream B camp (UpB) and Downstream B camp (DnB) on ice stream B are also shown. The boxes outlined in black are the region of radar coverage of *Retzlaff et al.* [1993], used by *Alley et al.* [1994] to calculate a hydrologic potential map. The box outlined in white is the region displayed in Fig. 1, and includes the southern flank of Siple Dome, the Duckfoot and the grounding line..

tation of *Hughes*, 1975] or along the edge of Siple Dome near ice stream C [interpretation of *Shabtaie et al.*, 1987]. After further radar surveys, *Rose* [1979] termed ice stream C "an enigma"— radar clutter indicated open crevasses, but visual observation showed an unbroken surface. He suggested that ice stream C was formerly active but had stopped; open crevasses that had formed during the active stage were now buried but still visible with radar. He also suggested that ice stream C is building up in its upper reaches possibly preparatory to resumption of fast flow, and that ice stream B is "now ...regaining ...lost territory" in the region between the heads of ice streams B and C owing to quiescence of C.

Extensive airborne radar with ground control [*Shabtaie and Bentley*, 1987; *Shabtaie et al.*, 1988; *Retzlaff et al.*, 1993] produced more-accurate maps of the ice thickness and bed elevations. Ice stream C extends more than 400 km from the grounding line near Siple Dome to the "onset region". The lower section of the ice stream has stagnated while velocities farther upstream (east of 125°W) are $\sim$100 $m \cdot a^{-1}$ which, together with evidence of open crevasses and streaming flow from satellite images [*Hodge and Doppelhammer*, 1996; *Bell et al.*, 1998] suggests that the upper part of ice stream C may still be active. This is supported by recent interferometric RADARSAT data that indicate velocities up to 70 $m \cdot a^{-1}$ in the source area [*Joughin et al.*, 1999]. Further, Joughin and others confirm and extend the identification by *Hodge and Doppelhammer* [1996] of two tributaries (C1 and C2) feeding the main stream. The northern tributary (C2), previously referred to as the limb [*Alley et al.*, 1994], has buried shear margins similar to those of the downglacier part of ice stream C, referred to as the trunk. The southern tributary, which was located using ground GPS measurements [*Anandakrishnan et al.*, 1998] and satellite interferometry, does not appear to have a continuous, buried shear margin (though individual buried crevasses do occur along the margins).

The tributaries come together just above the slow-flowing trunk near 132°W (about 50 km upstream from the old UpC camp—Plate 1) resulting in thickening there of 0.49 $m \cdot a^{-1}$ [*Joughin et al.*, 1999]. This contrasts sharply with neighboring ice stream B, which is fast flowing along its entire length (u $\sim$400–800 $m \cdot a^{-1}$) and thinning rapidly (estimated at 0.06–0.12 $m \cdot a^{-1}$ [*Whillans and Bindschadler*, 1988; *Shabtaie and Bentley*, 1987, *Whillans et al.*, this volume]). This inverted flow pattern of fast-flowing tributaries converging on a slow-flowing trunk demonstrates that ice stream C is not in a steady-state configuration but it is still responding to the recent stagnation of the trunk.

The stagnation of the trunk of ice stream C has been variously interpreted as:

- evidence of stability in the ice sheet system: some negative feedback slowed ice motion that had become too vigorous, perhaps related to surging behavior in which fast and slow flow alternate because of internal dynamics [e.g. *Rose*, 1979; *Radok et al.*, 1987; *Retzlaff and Bentley*, 1993]. On the other hand, the rapid change might indicate that other such rapid changes, including ones that might destabilize the ice sheet, are possible [*Alley and Whillans*, 1991],
- a response to ongoing drawdown of the ice sheet [*Alley et al.*, 1994; *Anandakrishnan and Alley*, 1997b], possibly leading to its collapse [*Bindschadler*, 1997].

Understanding the change from fast to slow flow of ice stream C may be central to assessing the present stability of the WAIS.

2. TIME OF SLOW-DOWN

There is little question that ice stream C was once active and that the trunk region has now slowed greatly. Surface-based radar profiles show folding and deformation of the internal layers characteristic of fast ice flow [*Jacobel et al.*, 1993]. Airborne radar profiles reveal characteristic ice stream "clutter" (a prolongation of the surface echo due to buried crevasses and surface inhomogeneities) within the body of the ice stream [*Shabtaie and Bentley*, 1987]. Particularly strong clutter is characteristic of the boundaries of ice streams where high lateral shear strain rates produce a chaotic crevassed zone along the margins [*Raymond et al.*, this volume]. In contrast, the slower-flowing interstream ridges typically produce a short surface echo in the radargrams because surface inhomogeneities are primarily sastrugi rather than crevassing. The presence of clutter and buried paleo shear-margins are evidence that ice stream C was once similar to neighboring ice streams B and D, and flowed at speeds in excess of 120 $m \cdot a^{-1}$(necessary to maintain an active shear margin [*Scambos and Bindschadler*, 1993]).

Ground-based, high frequency radar has been used to measure the depth to the tops of buried crevasses on ice stream C [*Retzlaff and Bentley*, 1993; *Bentley et al.*, in press]. These measurements, when combined with age-depth profiles determined from snowpits and cores analysed for accumulation rate [*Whillans et al.*, 1987; *Whillans and Bindschadler*, 1988], yield an estimate of the time of slow-down.

Five radar profiles across the southern margin of the ice stream during the 1988-89 Antarctic field season showed depths to the tops of the shallowest buried crevasses ranging from ~ 7 m (near the upstream end at $120° W$) to ~ 20 m farther downstream [*Retzlaff and Bentley*, 1993]. Using estimates of the accumulation rate, they inferred the slow-down of the trunk region occurred 140 ± 30 a BP, and a more recent slow-down farther upstream. Recent work across the northern margin showed a similar pattern [BE Smith, pers. comm., 1998]. West of $125° W$, the crevasse tops are 20 to 30 m below the surface. Just east of $125° W$ depths are less than 10 m and open crevasses were observed farther to the east. These newer measurements (along with accumulation rate and density data) indicate that the trunk of the ice stream stagnated nearly synchronously at 150 ± 30 a BP.

We note that age-depth profiles are critical for determining the time of shut-down, and published measurements of the accumulation rate and density-depth profiles in the region are sparse [*Whillans and Bindschadler*, 1988]. Furthermore, analysis of the spacing between radar-detected internal layers indicates accumulation may vary by up to 30% within 10 km [BE Smith, pers. comm., 1998]. Nevertheless, the emerging picture is that the lower section of ice stream C stagnated nearly sychronously about 150 a BP, while the upper section (east of $125° W$) is still actively forming crevasses.

3. THERMAL CONDITIONS AT THE BED

It has been clear from the earliest work that the bed of ice stream C remains wet in most places. The pseudo ice shelves of *Robin et al.* [1970] were recognized in part based on the especially bright basal reflections in radar, indicating a wet bed. *Rose* [1979], *Shabtaie and Bentley* [1987] and *Shabtaie et al.* [1987] extended this work, demonstrating a wet bed with predominantly fresh rather than salt water. *Shabtaie et al.* [1987] suggested the possibility of centimeters-thick water or thicker in some places, with uncertainties caused by the numerous corrections required to estimate reflection coefficients from returned radar power. Recent reanalyses of these airborne radar data, which were collected during the 1987-88 field season and cover the downstream portions of ice streams B and C, show pronounced contrasts in the reflection-strength between the ice streams and ridge BC [*Bentley et al.*, 1998]. Reflection strengths are interpreted to confirm the idea that ice stream C is not frozen at the bed.

Seismic studies by *Atre and Bentley* [1993] similarly demand basal melting to explain the low acoustic impedance of the bed (any debris-rich frozen material would have given significantly higher acoustic impedances than observed in many places). The degree of basal lubrication demonstrated by *Anandakrishnan and Bentley* [1993] also was inconsistent with any model of widespread freezing to rigid basal material.

These observations suggest that although a thawed bed may be necessary for fast flow, it is not sufficient. The role of basal water has figured prominently in speculations on mechanisms for the slow-down of ice stream C. *Rose* [1979] calculated that either fast ice motion or high geothermal fluxes were needed to maintain the bed at the pressure melting point in steady state. *Shabtaie et al.* [1987] also noted that the surface of the trunk of ice stream C is terraced, and in places, the longitudinal slope reverses. They pointed out that using a one-dimensional model, as they did, instead of a more-realistic two-dimensional model makes it difficult to accurately model the basal hydraulic potential along the ice stream, but noted that the reversals in basal water flow may contribute to the observed spatial variability in the basal reflection coefficient.

4. HYPOTHESES FOR SLOW-DOWN

Several hypotheses for the shut-down of the lower trunk of ice stream C have been proposed:

Surging: The ice streams exhibit periodic dynamic oscillation [*Rose*, 1979; *Hughes*, 1975; *Lingle and Brown*, 1987] akin to that of some mountain glaciers [cf. *Kamb et al.*, 1985; *Clarke*, 1987a]. Questions are raised about this by the difficulty of modeling surges in the ice streams [*Radok et al.*, 1987]. Also, surging mountain glaciers spend most of their time in the slow-flow mode, whereas most of the Siple Coast ice streams are in fast-flow mode [*Clarke*, 1987b].

Surging via basal water feedbacks: In this model, ice flow speeds up until basal-water generation becomes sufficiently large that the water experiences the *Walder* [1982] instability, channelizes, lowers basal water pressures, and so slows or stops the ice [*Retzlaff and Bentley*, 1993, *WB Kamb & HE Engelhardt*, presented at Chapman Conference, Orono, ME, 1998]. Without rapid motion, the basal water channels could not be maintained, and basal water pressures eventually would rise and allow resumption of fast flow; interactions involving repeated capture of drainage basins by neighboring ice streams are suggested. The subsequent work of *Walder and Fowler* [1994] raises questions about the viability of this mechanism; on an unconsolidated sediment bed such as that indicated for ice stream C based on seismic results [*Atre and Bentley*, 1993] and direct coring [WB Kamb, WAIS Meeting, Sept. 1999], water channels are expected to show increasing water pressure with increasing flux. Emerging evidence that side drag is important or dominant in restraining active ice streams because basal lubrication is exceedingly efficient [*Echelmeyer et al.*, 1994; *Harrison et al.*, 1998; *Raymond*, 1996; *Whillans et al.*, 1993] also leads to questions of whether sufficient water could have been generated from the viscous dissipation of fast flow to allow the *Walder* [1982] instability. However, much of the evidence for dominant side drag and minimal bed drag is from ice stream B, which is known to have a smoother bed, hence potentially better basal lubrication, than the other Siple Coast ice streams [*Jankowski and Drewry*, 1981].

Loss of lubricating till: Several lines of evidence indicate that the soft tills known to exist beneath ice streams B and C are important in the rapid ice motion, through some combination of till deformation, burying of bedrock bumps, or allowing ploughing of controlling-obstacle-size bumps [e.g. *Alley et al.*, 1987; *Blankenship et al.*, 1987; *Brown et al.*, 1987; *Kamb and Engelhardt*, 1991]. Reduction or loss of that lubricating layer might slow or stop the ice motion [*Retzlaff and Bentley*, 1993]. The persistence of a widespread soft-sediment layer beneath the now-stagnant trunk of ice stream C [*Atre and Bentley*, 1993; *Anandakrishnan and Bentley*, 1993; *Anandakrishnan and Alley*, 1994, 1997b] argues against such a model, however. Similarly, evidence of a deep sedimentary reservoir upstream [*Anandakrishnan et al.*, 1998] suggests that the erosional source still exists. Finite-element model experiments by *Fastook* [1987] to validate an ice-piracy scenario versus a loss-of-till scenario suggest that the latter is more likely than the former.

Ice-shelf backstress: Assuming that ice stream subglacial sediments deform, and depending on the velocity-depth profile in the sediment [e.g. *Alley*, 1989], the rate of deposition could range from very small to quite large. It has been suggested that the lightly grounded "ice plain" at the mouth of ice stream B is the result of such deposition [*Alley*, 1989]. *Thomas et al.* [1988] suggested that deposition at the mouth of the ice stream may have increased grounding and backstress, stopping streaming flow. Ongoing grounding-line retreat (measured at about $30\,\mathrm{m\cdot a^{-1}}$ between 1974 and 1984 [*Thomas et al.*, 1988]) in response to this stoppage might someday allow resumption of rapid flow. A potential difficulty with this hypothesis is that ice stream B maintains vigorous flow despite a large ice plain and despite the presence of Crary Ice Rise that provides significant restraint on flow [*MacAyeal et al.*, 1987, 1989]. Further, the results of *Anandakrishnan and Alley* [1997b] show that the grounding-line region of ice stream C provides little restraint on ice flow.

Ice piracy: "Piracy" is a concept borrowed from fluvial geomorphology defined as "the natural diversion of the headwaters of one stream into the channel of another stream having greater erosional activity and flowing at a lower level" [*Bates and Jackson*, 1980]. By analogy, if ice stream B in some fashion became better lubricated than ice stream C, hence faster flowing, the upglacier regions of ice stream B might thin, and ice would flow down the surface slope from the catchment of ice stream C into ice stream B. The surface of ice stream C then might flatten as ice flowing from its upper reaches was not replaced from the catchment, leading to stoppage. The biggest problem with this model is that the catchment of ice stream C does not appear to be feeding ice to ice stream B [*Shabtaie et al.*, 1988; *Retzlaff et al.*, 1993], and it is possible in fact that ice stream C has pirated the catchment of ice stream B [*Joughin et al.*, 1999].

Water piracy: Because the hydraulic potential of subglacial water is affected by bed elevation as well as ice pressure (and other factors such as degree of channelization of flow), water and ice flow need not be tightly coupled. The surface slope is about ten times more effective than the bed slope in controlling water flow direction, but steep transverse bedrock and flat ice surface slopes identified by *Rose* [1979], *Shabtaie and Bentley* [1987], and *Retzlaff et al.* [1993]

in the upglacier reaches of ice streams B and C suggest the possibility of water piracy, with lubricating water from the catchment of ice stream C diverted to ice stream B. A map of hydrological potential made from previously collected radar data [*Retzlaff et al.*, 1993] to test this idea [*Alley et al.*, 1994] shows that such water diversion probably is occurring, although the error bars include the small possibility that it is not. We favor this hypothesis as the cause of the shut-down of ice stream C [*Alley et al.*, 1994; *Anandakrishnan and Alley*, 1997a], though several difficulties remain [S Price and others, WAIS Meeting, Sept. 1999; see Sec. 6.1].

Thermal processes: Thermal processes have been suggested as controls on alternating fast and slow ice flow [*MacAyeal*, 1993b; *Payne*, 1995]. Thick ice traps geothermal heat and favors a thawed bed. However, rapid flow can bring cold ice near the bed through horizontal and vertical advection, and can thin ice so that the cold surface is closer to the bed. An oscillation has been modeled for the former ice sheet in Hudson Bay/Hudson Strait, linked to the Heinrich events of the North Atlantic [*MacAyeal*, 1993a, b]. A similar scenario is suggested for ice stream C [S Price, WAIS meeting, Sept. 1999; A Payne; WB Kamb & HE Engelhardt, Chapman Conference, Orono, ME, Sept. 1998]

Sticky spots: There is strong evidence for large spatial variation in basal drag of ice streams [see *Alley*, 1993; *MacAyeal et al.*, 1995]. If a localized region has a high basal shear stress, the water pressure in a distributed, connected basal water system will be reduced in that region. Thus, under an ice stream with sufficient water supply, sticky-spots would remain well lubricated and would not restrain flow. If the water supply were eliminated, as we hypothesize for ice stream C under the water-diversion scenario, these sticky-spots would play a significant role in restraining flow. The sticky spots would also be the first sites to freeze on in a thermal-shut-down scenario.

5. OBSERVATIONS AND DATA

5.1. Current flow pattern of ice stream C

Though the flow speeds of the trunk are low (comparable to ice-sheet or inter-ice-stream flow speeds), the ice stream is not frozen to its bed as discussed in Sec. 3. The trunk is flowing parallel to the main axis of the ice stream, towards the Ross Ice Shelf, though there appears to be locally divergent flow at UpC [HE Engelhardt, WAIS meeting, Sept. 1999]. The flow pattern of the tributaries feeding the ice stream is also complex. The southern tributary is located above a deep low-density sedimentary basin as determined by seismic, aeromagnetic, and gravity measurements [*Anandakrishnan et al.*, 1998; *Bell et al.*, 1998]. The velocities along a transverse line that crosses the southern margin of this tributary change from $12\,\mathrm{m{\cdot}a^{-1}}$ to over $60\,\mathrm{m{\cdot}a^{-1}}$ over a distance of 4.5 km, and distinct flow bands are visible in satellite imagery. This basin appears to control the position of the margin [*Anandakrishnan et al.*, 1998] and of the onset position of the ice stream [*Bell et al.*, 1998]. In addition, this basin could be a source of sediments for the ice stream subglacial till layer.

The northern tributary extends farther inland, following the Bentley Subglacial Trough. The onset location picked by *Hodge and Doppelhammer* [1996] from surface features does not correspond to a distinct change in flow speeds. Some component of the fast flow in the Bentley Trench is likely due to the thick ice there; it is unknown whether sediments are present there or not. The two tributaries coalesce upstream of the trunk (about 50 km upstream of station UpC) [*Joughin et al.*, 1999].

5.2. Shut-down of Siple ice stream and Duckfoot

At the eastern tip of Siple Dome (the ridge between ice streams C and D) ice stream C cuts across the older relict "Siple ice stream" [*Jacobel et al.*, 1996]. The age of stagnation for Siple ice stream is 420 to 470 a BP [*Conway and Gades*, in press, BE Smith, Chapman Conference, Orono, ME, Sept. 1998]. The cause of this shut-down is unknown, but we suggest that a similar water diversion occurred here as has been hypothesized for ice stream C. Because of the longer time since shut-down, the surface topography (and consequently the basal hydrologic potential) will be substantially different from those that existed four centuries ago.

The north margin of ice stream C runs along the flank of Siple Dome where satellite imagery (Figure 1) reveals a splayed pattern of margin scars and flowbands called the "Duckfoot" [RW Jacobel, TA Scambos, NA Nereson, and CF Raymond, Changes in the margin of ice stream C, *J. Glac.*, in review]. Though the underlying cause is not known, it appears that the north margin of C shifted inward with an accompanying change in flow direction. As it did so, ice from the area between the old and new mar-

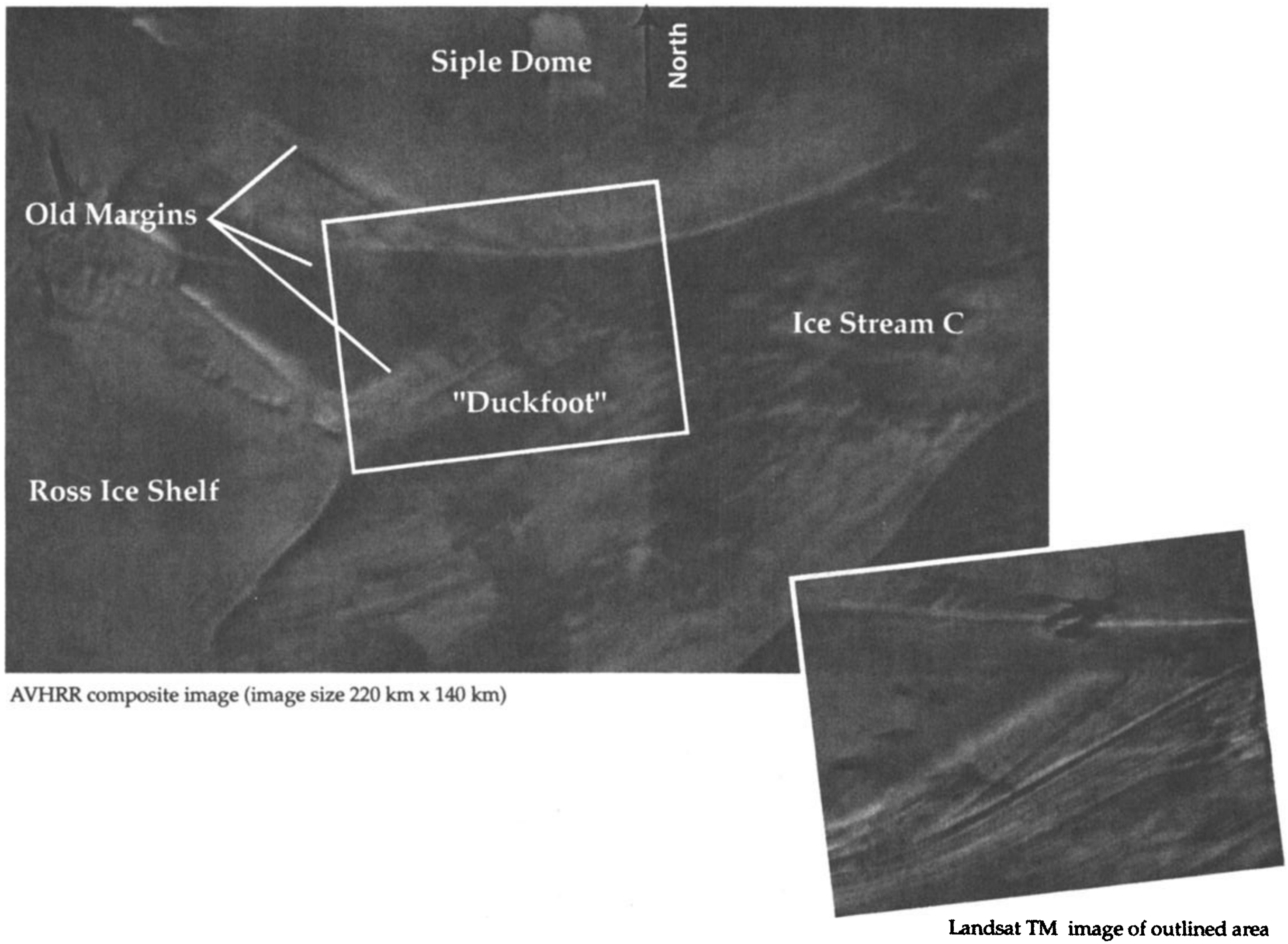

Fig. 1. An AVHRR (Advanced Very High Resolution Radiometer) satellite image of Siple Dome and inset Landsat image of the Duckfoot area. The edges of the main image are outlined in Plate 1. From Jacobel and others, in review.

gins did not stagnate immediately, and parts of it were sheared and folded by flow along the new direction of motion [Fig. 3 in *Jacobel et al.*, 1996] before stagnation. The lack of any substantial difference in the surface texture of the two regions suggests that the margin shift did not appreciably precede the shut-down of C, but it may be inferred that the outer margin is older because of the clear evidence of the inner one, and the presence of shearing between the two. Presumably all flow traces of the Duckfoot would have been transformed to normal flow striping if that area were part of an active ice stream for any significant time.

The longitudinally compressive strain rate observed in ice motion studies across the lower portion of the flank of Siple Dome [*Nereson*, in press] implies that this area must be thickening considerably, probably as a direct result of the shut-down and the thickening of ice stream C. The timing of shut-down of an ice stream may be determined by examining the rate of thickening of the adjacent ridge [*Nereson*, in press, Jacobel and others, in review]. According to this analysis, a wave of thickening which travels faster than the ice caused small topographic features associated with the former margin of an ice stream to be lifted onto the flank of the ridge. As the dome reaches a new steady state, ice flow carries the feature down slope. Models of this process suggest that the outermost Duckfoot scar is experiencing the beginnings of this rise. The pattern of thickening (inferred from the ice flow field for the south flank of Siple Dome) is consistent with a margin shut-down approximately 300 to 500 years before present.

This scenario proposes that the surface features on the north side of ice stream C are the result of a widening stagnation of a triangular-shaped area of ice just inside the older margin. The root causes of this stagnation are not yet clear, but some possibilities are [see Jacobel and others, in review]: (1) a general reduction in water pressure under ice stream C which caused the largest proportional increase in effective normal stress at the bed in this marginal zone because the ice was thinnest there; (2) reduced velocity in ice stream C, thus reducing the drag on this margin from the central parts of the ice stream; (3) onset of freezing associated with a general thinning of ice stream C which would have had the strongest and most rapid effect where the ice is thin; (4) water piracy that swept from the north to the south resulting in progressive shut-down.

5.3. Basal Environment of ice streams B & C

Ice stream B is thawed at the bed and fast-flowing along its entire length (see *Whillans et al.*, this volume). Much has been learned about the bed of ice stream B from seismic imaging, radar work, and direct drilling performed over the last decade. These experiments show that there is high water pressure at the bed and that there exists a low-strength subglacial till layer. The velocity is partitioned in some still-debated way between deformation in the till layer and sliding of ice over its base on the water layer [*Engelhardt and Kamb*, 1998, RB Alley, Continuity comes first: Recent progress in understanding subglacial deformation, Proc. of the Intl. Conf. on Deformation of Glacial Materials, London, Sept. 1999, in review].

Seismic imaging of ice stream B revealed a heterogenous bed. In particular, the till layer pinches out (or nearly so, within the resolution of the seismic experiment) along a longitudinal ridge. This region of higher strength material (and possibly other, similar regions where the till is absent) would present an obstacle to flow.

The bed of ice stream B appears to provide little resistance to flow, with lateral drag on the margins supporting most of the driving stress for ice flow [*Whillans and van der Veen*, 1997], though *Raymond et al.*, [this volume] suggest that ice streams D & E have significant support from the bed. Of the fraction of driving stress supported by the bed of ice stream B, most is supported on the till rather than on the rare sticky spots [*Alley*, 1993] despite the weakness of the till (yield strength of 2 kPa [*Kamb*, 1991]). The sticky spots are well-lubricated and at UpB camp, it is estimated that $<13\%$ of the basal shear force is supported by sticky spots.

This calculation is bolstered by the observation of rare basal microearthquake activity at UpB and no basal seismic activity at DnB [*Blankenship et al.*, 1987; *Anandakrishnan and Bentley*, 1993]. Seismic monitoring of ice stream B at UpB camp (1985–86) showed that the basal microearthquake activity was low (but significantly, non-zero). At UpB, six events were recorded that emanated from the bed of the ice (within the hypocentral depth determination error of ± 15 m) in 85 hours of monitoring [*Blankenship et al.*, 1987]. All the events were coincident in space (within the epicentral location ellipse of

± 10 m) and occurred within a half-hour period. The events were low-angle thrust faults with slip in the direction of ice flow and of very small magnitude (seismic moment $\approx 10^6$ N $\cdot$ m). The interpretation of these events is of a transient increase in basal shear force on a local, more-competent portion of the bed, followed by fracture and slip [*Anandakrishnan and Bentley*, 1993].

We estimate that the stress drop for these events was approximately 10 kPa, which is estimated to equal or exceed the average basal shear stress of ice stream B. As stress drop is usually only a fraction of the total applied stress on the fault (between 1 and 10%), it is likely that the sticky spot was supporting the driving stress from some larger portion of the bed than simply that of the area of the sticky spot. That is, much or all the basal shear stress from that larger area is concentrated on the sticky spot and as a consequence the material fails. We caution that estimates of fault-plane area, stress drop, and the fraction of applied stress are strongly slip-model dependent and therefore inaccurate. Nonetheless, the presence of repeated fracture at a single spot at the bed that is induced by the shearing force of the ice, is evidence of at least one sticky spot beneath ice stream B. Other sticky spots (estimated to cover 2-3% of the bed at UpB [*Rooney et al.*, 1987; *Rooney*, 1988]) remained non-seismic and presumably well-lubricated throughout the seismic monitoring experiment.

5.4. Microearthquakes along ice stream C

To determine bed characteristics in the trunk and in the tributaries we measured the rate of basal seismicity along the ice stream. The hypothesis was that well-lubricated beds have low seismicity, but poorly lubricated beds that are not mostly-frozen would have high seismicity.

Surprisingly, the trunk of ice stream C was highly active seismically, with tens to hundreds of basal thrust-fault events recorded per day [*Anandakrishnan and Bentley*, 1993]. These events preferentially occured and recurred on localized sticky spots of order 10 m linear dimension, separated by order 100–1000 m. Quakes beneath ice stream C triggered other quakes on adjacent sticky spots, to distances as great as 1.5 km, and with time delays indicating propagation of the disturbance at approximately 1.9 m$\cdot$s^{-1} [*Anandakrishnan and Alley*, 1994]. The microearthquakes were first observed in 1988 [*Anandakrishnan*, 1990; *Anandakrishnan and Bentley*, 1993] and remeasured in 1995 and 1996 [*Anandakrishnan and Alley*, 1997a].

Seismometers were deployed at approximately 90 km separations along the length of ice stream C, from just above the grounding line to above the onset of streaming flow. The sites are labeled by their distance from the grounding line (that is, site 10 is 10 km upstream of the grounding line, and so on; Plate 1). The rate of basal seismicity R (number of events per day) was low for the two sites on ice stream B (UpB and DnB), and for sites Km 482 and Km 432 (in the catchment of ice stream C, and in the uppermost part of ice stream C, respectively). There was a marked downglacier increase in seismicity between Km 432 and Km 354 and seismicity remained high from Km 354 down to the array closest to the grounding line at Km 10 (Fig. 2). Flow velocities were low on the ice sheet (Km 482); the ice flowed faster in the upper reaches of ice stream C (Km 432 to Km 354: $30 < u < 60\,\mathrm{m\cdot a^{-1}}$) but nearly stagnated somewhere between Km 354 and Km 252 ($u < 10\,\mathrm{m\cdot a^{-1}}$ [*Anandakrishnan and Alley*, 1997a; *Whillans and van der Veen*, 1993]). Thus the pattern is clear: on the ice streams, low velocities are associated with high seismicity and vice-versa. The anomaly in this pattern is Km 354, which had a relatively high velocity but also had high seismicity. This site appears to be transitional between streaming and non-streaming ice and exhibits some of the qualities of each.

5.5. Thawed bed

The bed is thawed and exceptionally well lubricated by a soft till almost everywhere [cf. *Atre and Bentley*, 1993; *Bentley et al.*, 1998; *Anandakrishnan and Alley*, 1997b, *WB Kamb & HE Engelhardt*, Chapman Conference, Orono, ME, Sept. 1998], but with localized, poorly lubricated regions. The till is unfrozen and contains water at high pore pressures, but whether a distributed, connected water system exists under C (as exists under ice stream B [*Engelhardt and Kamb*, 1997]) is unknown. The observation of high seismicity from sticky spots at UpC over an extended period of time (1988 to 1996) suggests that there is not sufficient free water to flow down the local hydrologic potential gradient and lubricate the sticky spots. We measured different sticky spots in 1988 and in 1996 because the arrays were separated by about 10 km, but the presence of high-friction regions at the bed over an eight year period suggests that the basal water system is poorly developed.

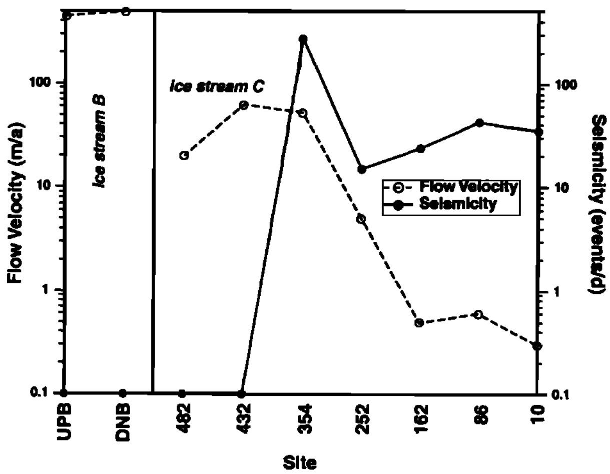

Fig. 2. Plot of ice-flow velocity u (m · a $^{-1}$) and seismicity R (day $^{-1}$) at the different array locations (from *Anandakrishnan and Alley* [1997a]). Note that the values of $R = 0$ are plotted at $R = 0.1$ because of the log scale.

The presence of a meters-thick layer of till under a different part of the ice stream was inferred by the transmission of tidal forcings upstream from the grounding line [*Anandakrishnan and Alley*, 1997a]. They discovered that the rate of basal seismicity was related to the tide beneath the Ross Ice Shelf. The seismicity at the grounding line is in phase with the tide and peaks at local low tide. The peak seismicity at a location 80 km inland from the grounding line (and on the ice stream) lagged the low tide by 13 hours. They successfully modeled the ice-stream system as an elastic beam resting on a viscous substrate and showed that a linear-viscous till was consistent with the measurements but a non-Newtonian rheology (high exponent in stress-strain relationship $\dot{\epsilon} \propto \tau^p$, $p \sim 100$) did not fit the data. We note that it is possible that till will behave as a high-exponent plastic material under high strain rates (and large total strain) as occurs under ice stream B, but will be linear-viscous under the lower strain rates at the base of nearly stagnant ice stream C [*Iverson et al.*, 1998, Alley, in review, see Sec. 5.3].

6. DISCUSSION

The near-stoppage of the trunk of ice stream C is clearly of considerable interest, and in Sec. 4 we identified 8 hypotheses for this shut-down. Of these, we favor water piracy. We believe that five are rendered highly unlikely by the data: there is as yet no evidence for classical surging behavior, soft till remains beneath the ice and is modeled to prevent the *Walder* [1982] water-system instability, ice plains are too well lubricated to be significant in the force budget, and ice piracy has not yet occurred. Thermal processes and sticky spots remain potential explanations, and we are intrigued by the possibility that water piracy, sticky spots, and thermal processes have all contributed to the changes in ice stream C, as discussed next.

6.1. *Water diversion*

As noted in Sec. 4 ("water piracy" hypothesis), *Alley et al.* [1994] hypothesized that ice stream C slowed because of loss of water that lubricates sticky spots where till is thin or absent. A hydrological potential map of ice streams B & C made to test this hypothesis indicated that water from the catchment area of ice stream C is being diverted to ice stream B. Despite the high data quality and tight flight-line spacing, the data errors might allow limited or zero water diversion, although strong water diversion is clearly the better interpretation.

The site of probable water diversion is near where the upglacier tributaries join the downglacier trunk (Plate 1). *Alley et al.* [1994] thus hypothesized that the tributaries remain active but aseismic because their sticky spots are water-lubricated, while the trunk nearly stopped and is seismically active on the sticky spots. The hydrologic potential is ten times as strongly affected by surface topography as by bed topography. Thus, water will tend to flow in the direction of the ice surface slope. However, with the low average surface slopes of the ice stream ($\alpha \sim 0.001$), the bed topography under the hypothesized diversion area becomes significant in affecting basal water flow. The large transverse bed slopes in the diversion zone would not control the basal water flow if the surface of the overlying ice were steep, as is common with ice-sheet ice. Under those conditions, which *Alley et al.* [1994] suggest existed in the past, the basal water would flow in the direction of the ice-sheet surface slope. Thus water from the catchment of ice stream C was directed towards the ice stream even in the presence of large transverse bed slopes. With Holocene warming and the subsequent drawdown of the ice sheet, ice stream C grew headward and the low ice-stream surface slopes impinged on this region of transverse bed slopes [*Alley and Whillans*, 1991]. This allowed the transverse bed slopes to dominate the hydrologic potential and the water from the catchment of ice stream C flowed towards ice stream B. The loss of this water layer was hypothesized to result in a loss of lubrication of sticky spots, an increased basal friction, and a stagnation of the ice stream below the water-diversion zone.

A bulge at the confluence of the tributaries is a result of the tributary ice "piling up" against a stagnant trunk as noted by *Alley et al.* [1994]. The profile of the confluence during the time of active trunk flow might have been concave up, preventing water diversion from the confluence area. Thus the suggestion of Price and others (WAIS meeting, Sept. 1999) is that the hydrologic potential that diverts basal water from ice stream C to B is a recent phenomenon and not the cause of the stagnation of the trunk [cf. *Alley et al.*, 1994]. We note, however that transverse ice stream surface elevation profiles are generally concave up in the downstream portions, but flat to convex up in the upper portions [see *Shabtaie et al.*, 1987, Fig. 3a]. Further, we note that even partial diversion of basal water from, e.g., one side of the ice stream to the other side would be nearly as effective at stagnating the ice stream as full diversion from ice stream C to ice stream B. Thus, a flattening of surface profile (a progressive change from longitudinally convex up to concave up during head migration of the ice stream) over a region of steep transverse bed slopes could funnel water to the southern half of ice stream C, possibly resulting in rapid stagnation. The present-day observation of water diversion to ice stream B would then be a later phenomenon. More work is needed on this hypothesis.

6.2. Thermal Processes

Price and others (WAIS meeting, Sept. 1999) argued in favor of a thermal model for the shut-down of ice stream C [cf. *MacAyeal*, 1993a; *Payne*, 1995]. Thermal surging is complicated by basal water transport in ice-contact systems or in subglacial till [*Alley and MacAyeal*, 1994]. Basal water can be considered to be stored thermal energy from beneath ice upglacier, and provides a heat source to any region where freezing is initiated through its latent heat. For deforming tills, the effectiveness of this heat source will depend in large part on the existence or absence of sticky spots of thin or absent till–without such sticky spots, the ice cannot freeze to bedrock until water in till pores is frozen, but sticky spots might allow freeze-on more quickly. Freeze-on in the presence of an active ice-contact water system likely requires that most or all of the water flux be frozen before the ice can freeze to its substrate, which could greatly suppress freezing-on for significant water fluxes. If water is not supplied from upglacier, freeze-on may contribute to consolidation of subglacial tills, strengthening the bed and resisting rapid ice motion [*Tulaczyk et al.*, in press]. The persistence of soft till beneath ice stream C (Section 5.5) demonstrates that this process has not proceeded far for much of the ice stream, but the process may have contributed to generation of the observed sticky spots, perhaps in regions of previously thin till over bedrock bumps.

Here, the water-piracy hypothesis and the thermal hypothesis may be complementary. Water piracy would have caused a significant loss of heat as well as lubrication to ice stream C. The stoppage of ice stream C may result from loss of lubrication of sticky spots, from incipient freeze-on to sticky spots (the intervening till remains soft), or from some combination of these end-members.

7. CONCLUSION

The rapid flow of Siple-coast-type ice streams depends on the bed of the ice presenting little or no resistance. Variability in bed properties such as bedrock knobs or non-uniform distribution of deformable till could present a significant resistance to flow unless these sticky spots are lubricated by water that decouples the ice. The Siple ice streams (with the exception of ice stream C) appear to receive sufficient water from their catchments and produce more water by melting of the bed due to fast sliding [*Engelhardt and Kamb*, 1997]. This basal water system efficiently nullifies the restraining forces of the sticky spots and allows the ice streams to maintain the high flow speeds observed. This thin water layer is hypothesized to lubricate sticky-spots under ice stream B and under the upstream portion of ice stream C (above the diversion zone). The sticky spots are regions of higher basal shear stress than their surroundings and are modeled to have lower water pressures. Thus, the well-connected basal water system can deliver lubricating water to the sticky spot, and reduce shear stress in a stable negative feedback mechanism.

We hypothesize that similar conditions currently exist under the upstream part of ice stream C, and existed under all of ice stream C as recently as 150 ± 30 a BP. We suggest that at that time, the headward growth of the ice stream brought low ice stream surface slopes over high transverse bed slopes resulting in a diversion of catchment water from ice stream C to ice stream B. As a consequence, the sticky spots were starved of lubricating water and could, and did, exert a restraining force on the ice stream. As a result, flow speeds of the lower part of the ice stream are less than $10\,\mathrm{m\cdot a^{-1}}$, and the ice at UpC is thickening at a rapid rate of up to $0.49\,\mathrm{m\cdot a^{-1}}$ [*Joughin et al.*, 1999]. An alternative suggestion is that the trunk freeze-on and stagnation occured first and the water diversion at the confluence is a consequence of that stagnation. A possible synthesis of these two primary hypotheses is that the water diversion occurred farther upstream (where transverse ice stream profiles might have been convex up even prior to stagnation). If so, we speculate that the regions downstream (including both the bulge and the trunk) were starved of water, but the trunk stagnated first and that the wave of stagnation will proceed upstream as more sticky-spots freeze on.

The mass balance of the combined ice stream B & C system (ice stream and catchment) appears to indicate slow thinning according to the best estimates available [*Shabtaie and Bentley*, 1987; *Shabtaie et al.*, 1988], though the few available point measurements show large spatial variability [e.g *Hamilton et al.*, 1998; *Whillans and Bindschadler*, 1988]. If the shutdown of ice stream C were a stabilizing influence, one might expect that C would be thickening and B remain in balance, resulting in a net thickening of the combined system. We suggest that the observed net thinning of ice stream B is possibly due to the extra basal lubrication provided by the water diversion from beneath C. Thus, the strong negative balance of ice stream B (as compared to approximately zero balance of ice streams D, E, and F [*Shabtaie and Bentley*, 1987]) is connected to general headward extension of the ice streams that resulted in the triggering of the shut-down of C.

The shut-down of ice stream C then is not an inherent feedback mechnism stabilizing the ice sheet, but a consequence of a particular combination of strong transverse bed slope in the onset region of the ice stream. To understand and predict the behavior of the other ice streams in the presence of ongoing headward migration, detailed knowledge of the basal environment (both topography and geology) is required.

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