Ice Stream C slowdown is not stabilizing West Antarctic Ice Sheet

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Abstract

Changes in the flow of ice stream C likely indicate a continuing slow drawdown of the Siple Coast of West Antarctica rather than a stabilizing feedback. The downglacier part of ice stream C, West Antarctica largely stagnated over the last few centuries, while upglacier regions continue to flow vigorously. 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. The soft till slows the transmission of ice-flow changes caused by microearthquakes or by tide-height in the Ross Sea, suggestive of viscous behavior. Near the transition from fast-moving well-lubricated ice to slow-moving ice with basal seismicity, 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, and consistent with the inland water continuing to lubricate fast ice flow, but now concentrated in ice stream B.

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1 Introduction

The West Antarctic Ice Sheet (WAIS) is strongly out of balance (defined as input precipitation minus outflow) along parts of the Siple Coast. Ice streams flowing from the interior ice sheet to the floating Ross Ice Shelf account for a large fraction of total outflow, and the outflow volume increases with the ice stream flow speed. The recent slowdown of ice stream C is therefore central to assesments of the stability of the West Antarctic. This stagnation has been variously interpreted as evidence of stabilizing feedbacks in the ice sheet flow system, or of an accidental response to ongoing drawdown possibly leading to collapse.

As recently as 140 years ago, ice stream C was flowing rapidly [*Rose*, 1979; *Retzlaff and Bentley*, 1993; *Shabtaie et al.*, 1987a]. The ice stream is currently slow-flowing along much of its length with flow speeds of $u < 10\,\mathrm{m \cdot a^{-1}}$ in parts and nowhere greater than $100\,\mathrm{m \cdot a^{-1}}$. This ice stream is thickening at the extremely rapid rate of 0.55 $\mathrm{m \cdot a^{-1}}$ at UpC [G Hamilton, Chapman Conference, Orono, ME, 1998]. Neighboring ice stream B is fast flowing ($u \sim 400$–$800\,\mathrm{m \cdot a^{-1}}$) and is thinning rapidly (estimated at 0.06–0.12 $\mathrm{m \cdot a^{-1}}$ [*Whillans and Bindschadler*, 1988; *Shabtaie and Bentley*, 1987]).

There is little question that ice stream C was once an active ice stream and that it then 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]. Particulary strong clutter is characteristic of the boundaries of ice streams where the strong lateral shear produces a chaotic zone of broken ice. The slower-flowing interstream ice and ice-sheet ice typically produce a short surface echo in the radargrams because surface inhomogeneities are primarily sastrugi rather than crevassing. The presence of clutter and of the marginal shear zone is evidence that ice stream C was once similar to neighboring ice streams B and D, and flowed at speeds in excess of $120\,\mathrm{m \cdot a^{-1}}$(required to maintain the chaotic ice of the shear margins; *Scambos and Bindschadler* [1993]). However, those marginal shear zones are now buried to a depth of between 7 and 20 m (see the detailed discussion of shutdown below). These depths to the crevasse tops suggest that the ice stream stagnated $140{\pm}30$ a BP [*Retzlaff and Bentley*, 1993] or $185{\pm}30$ a BP (from more recent accumulation rate data [Conway and others, ms. in prep]).

This is most easily interpreted as evidence of stability in the system: some negative feedback slowed ice motion that had become too vigorous, perhaps related to a general 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]. However, the mere existence of such a dramatic change can be interpreted to allow the possibility of further rapid changes that might destabilize the ice sheet [*Alley and Whillans*, 1991]. Furthermore, the near-stoppage of the trunk of C has been interpreted as an indication of ongoing drawdown of the ice sheet [*Alley et al.*, 1994b; *Anandakrishnan and Alley*, 1997b], possibly leading ultimately to the collapse of the ice sheet [*Bindschadler*, 1997].

2 Background

The early radar flying of *Robin et al.* [1970] identified ice stream B, and located several "pseudo ice shelves" either along ice stream C [interpretation of *Hughes*, 1975] or along the edge of Siple Dome near ice stream C [interpretation of *Shabtaie et al.*, 1987a]. 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 is a formerly active ice stream that stopped within the most recent 1000 years, and that the open crevasses formed during the active stage are present but buried. 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.

Shabtaie and Bentley [1987] used higher-resolution radar of the near-surface, together with new accumulation-rate data [*Whillans et al.*, 1987; *Whillans and Bindschadler*, 1988] to estimate that "the approximate date when the ice stream was last active" was about 250 years ago. This conclusion was later revised to "(ice stream C) ceased its activity less than about 250 years ago" [*Shabtaie et al.*, 1987a]. *Retzlaff and Bentley* [1993] then used short-pulse radar data from ground surveys to better constrain the near-stoppage of the ice stream along much of its trunk to about 130 years ago, synchronously within $\pm$ 30 years. Their farthest-upglacier survey, near where the trunk meets the limb of the ice stream, may have detected evidence of activity slightly more recently than 140 years ago, but the difference was not highly statistically significant.

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 (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.* [1987a] extended this work, demonstrating a wet bed with predominantly fresh rather than salt water, and suggesting the possibility of centimeters-thick water or thicker in some places [*Shabtaie et al.*, 1987a], although with uncertainties caused by the numerous corrections required to estimate reflection coefficients from returned radar power. The seismic studies by *Atre and Bentley* [1993] similarly demanded 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.

The role of the basal water has figured prominently in speculations on the lubrication and then stoppage of ice stream C. *Rose* [1979] calculated that fast ice motion or quite high geothermal fluxes were needed to maintain the bed at the pressure melting point in steady state. *Shabtaie et al.* [1987a] also noted that the surface of the trunk of ice stream C is terraced or shows slope reversals longitudinally [cf. *Robin et al.*, 1970], that a one-dimensional basal hydraulic potential model along the ice stream shows local reversals that may or may not exist in two-dimensional models, and that the possibility of reversals in basal water flow may be related to observed spatial variability in the basal reflection coefficient for radar. Attention has especially focused on the limb and near where the limb meets the trunk of ice stream B. *Rose* [1979] identified this as a region with anomalously flat ice-air surface slopes and steep bed slopes, and the later work of *Shabtaie et al.* [1987a] and *Retzlaff et al.* [1993] has verified and extended Rose's observations.

3 Hypotheses

Models of the shutdown of ice stream C include:

- **Surging** [*Rose*, 1979]**:** The ice exhibits some periodic dynamic oscillation akin to that of some mountain glaciers [cf. *Kamb et al.*, 1985]. 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.

- **Surging via basal water feedbacks** [*Retzlaff and Bentley*, 1993 *B Kamb & HE Engelhardt*, presented at Chapman Conf., Orono, ME, 1998]**:** 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. 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, 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 [*Whillans et al.*, 1993; *Echelmeyer et al.*, 1994; *Raymond*, 1996] 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 soft-sediment layer beneath the ice [*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 deformation of sub-ice-stream sediments occurs, deposition of this sediment at the grounding line should result. Depending on the velocity-depth profile in the sediment [e.g. *Alley*, 1988], the rate of deposition could range from tiny to quite large. It has been suggested that the lightly grounded "ice plain" region 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 the ice stream. 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. Possible difficulty with this model lies in the ability of ice stream B to maintain 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]. Also, the recent 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, and refers to "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], although continuation of ongoing trends [*Shabtaie et al.*, 1988] might cause such piracy in the future.

- **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 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 clearly 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.*, 1994b] 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 shutdown of ice stream C [*Alley et al.*, 1994b; *Anandakrishnan and Alley*, 1997a], though several difficulties remain.

4 Data

4.1 Shutdown of ice stream C

Ice stream C has two distinctly different flow patterns. The main body of the ice stream (the so-called trunk, the wider part of the ice stream west of 130° W) is slow flowing. The limb of the ice stream (the narrower, upstream part east of 130° W) is faster flowing at speeds that range from 30–60 $\mathrm{m \cdot a^{-1}}$ [*Anandakrishnan et al.*, 1998, *Conway et al.*, ms. in prep.], and possibly one location with speeds as high as $100\,\mathrm{m \cdot a^{-1}}$ [*Whillans and van der Veen*, 1993]. 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. 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. The flow pattern of the limb is also complex. The southern margin of the limb is located above a deep low-density sedimentary basin [*Anandakrishnan et al.*, 1998]. The velocity changes 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 boundaries of the nearly stagnant trunk of ice stream C have been described from satellite imagery [*Hodge and Doppelhammer*, 1996] and airborne radar [*Shabtaie and Bentley*, 1987]. High-resolution radar profiles show the pattern of internal layering (used to infer past patterns of flow and spatial variations in accumulation) and of particular interest, distinctive hyperbolae produced by buried crevasses [e.g. *Clarke and Bentley*, 1994].

Five radar profiles across the southern margin of ice stream C show depths to the tops of the shallowest buried crevasses ranging from $\sim 7\,\mathrm{m}$ (near the upstream end at $120^0\,W$) to $\sim 20\,\mathrm{m}$ farther down stream [*Retzlaff and Bentley*, 1993]. Using estimates of the accumulation rate, they inferred the slow down of the trunk region occurred $130 \pm 30\,\mathrm{a}$ BP, and a more recent slow down farther upstream. Recent work across the northern margin shows a similar pattern [*Bentley et al.*, in press; *Conway and Gades*, in press]. West of $125^\circ\,W$, the crevasses tops are 20 to 30 m below the surface. Just east of $125^\circ\,W$ depths are less than 10 m and open crevasses have been observed further to the east. Using an estimate of the local accumulation rate derived from snowpits and cores taken for beta analysis ($\sim 0.07\,\mathrm{m{\cdot}a^{-1}}$) gives a slightly older estimate of the stagnation time for the trunk region (180 ± 30 a BP). We caution that age-depth profiles are critical for determining the time of shutdown. Published measurements of the accumulation rate and density-depth profiles in the region are sparse [*Whillans and Bindschadler*, 1988] and analysis of the spacing between radar-detected layers in the region indicates accumulation may vary by up to 30% within 10 km [*Bentley et al.*, in press].

4.1.1 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). Recent crossings of this paleo-margin with both high and low frequency radar systems show disturbed layers and crevasse tops buried 40 to 50 m below the surface [*Bentley et al.*, in press; *Conway and Gades*, in press]. The corresponding age of stagnation for Siple ice stream is 420 to 470 a BP. The cause of this shutdown 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 shutdown, the surface topography (and consequently the basal hydrologic potential) will be substantially different than that which existed four centuries ago.

The north margin of ice stream C runs along the flank of Siple Dome where satellite imagery (Figure 3) reveals a splayed pattern of margin scars and flowbands called the Duckfoot [*Jacobel et al.*, 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 two margins did not stagnate immediately, and parts of it were sheared and folded by flow along the new direction of motion 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 shutdown of C, but it may be inferred that the outer margin is older because of the clear presence 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 compressive strain rate observed in ice motion studies across the lower portion of the dome flank [*Nereson*, 1998; *Jacobel et al.*, in review] implies that this area must be thickening considerably, probably as a direct result of the shutdown and the thickening of ice stream C. *Nereson et al.* [1998] studied what may be learned about the timing of shutdown of an ice stream from the rate of thickening of the adjacent ridge. According to this analysis, a wave of thickening which travels faster than the ice causes 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 shutdown 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 et al.*, in review]: (1) A general reduction in water pressure under ice stream C which causes the largest proportional increase in effective normal stress at the bed in this marginal zone because the ice is thin 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 will have the strongest and most rapid effect where the ice is thin. (4) That the water piracy swept from the north to the south resulting in progressive shutdown.

Recent radar studies (austral summer 1998-99) place the shutdown time of Siple ice stream at approximately 420 years BP [*Conway and Gades*, in press, *BE Smith et al.*, presented at the Chapman Conference, Orono, ME, Sept., 1998]. The approximate date for the shift in the north margin of ice stream C at the Duckfoot based on model studies is nearly the same. With the information currently available it is not possible to say more about a connection between these events. However, enhanced Landsat Imagery shows clearly that motion in the main trunk of ice stream C post dates the shutdown of Siple ice stream. This is evident from the tails of flow bands streaming north into Siple ice stream which have been sheared and truncated by more recent flow in ice stream C (Figure 3 in Jacobel et al., 1996).

4.2 Sticky Spots

There is strong evidence for large spatial variation in basal drag of ice streams [see *Alley*, 1993]. 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. The standard form of the basal hydrologic potential Φ is

$$\Phi = \rho_i g(z_s - z_b) + \rho_w g z_b \tag{1}$$

where g is the gravitational acceleration, ρ_i is the density of ice, ρ_w is the density of water, z_s is the ice-surface elevation, and z_b is the bed-elevation. Equation 1 accounts for the ice overburden pressure and the elevation of the subglacial water [*Paterson*, 1994]. *Weertman* [1972] suggested that the potential function that drives subglacial water flow at smaller scales, at which sticky spots would be important ($O(10^1$–$10^3)$ m)), is

$$\Phi_s = \Phi - K_n \tau_b \tag{2}$$

where τ_b is the basal shear stress, and K_n is a non-dimensional constant of $O(1)$. If a sticky spot were sufficiently sticky (i.e, if $\Delta\tau_b$ were large), then water would flow into it, increasing the lubrication and decreasing $\Delta\tau_b$. We define the local increase in basal shear stress at the sticky spot, above the average τ_b, as $\Delta\tau_b$.

4.3 Microearthquake activity—Ice streams B & C

Much work has been performed on neighbouring ice stream B. This ice stream is thawed at the bed and fast flowing along its entire length. Much has been learnt 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 question of whether this till layer is deforming or if the ice stream is sliding over its base on the water layer has not been settled.

Seismic imaging of the ice stream reveals 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 controlling the flow of ice [*Whillans and van der Veen*, 1997]. Even though the basal till is weak (yield strength of 2 kPa [*Kamb*, 1991]), *Alley* [1993] suggests that much of the basal shear is supported by the till. 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 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 appoximately 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. 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 well-lubricated throughout the seismic monitoring experiment.

Surprisingly, the trunk of ice stream C is highly active seismically, with tens to hundreds of basal thrust-fault events recorded per day [*Anandakrishnan and Bentley*, 1993]. Quakes preferentially occur and recur on localized sticky spots of order 10 m linear dimension, separated by order 100–1000 m. Quakes beneath ice stream C trigger 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 $1.9\,\mathrm{m{\cdot}s^{-1}}$ [*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].

4.4 Thawed bed

These results indicate that the bed is thawed and exceptionally well lubricated by a soft till almost everywhere [cf. *Atre and Bentley*, 1993; *Bentley et al.*, 1998 *B 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 does under ice stream B) 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 sticky-spot hydrologic potential gradient $\nabla\phi_s$ and lubricate the sticky spots.

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-Newtonial rheology (high exponent $p \sim 100$) 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.

4.5 Water diversion

We hypothesized that ice stream C slowed because of loss of water that lubricates sticky spots where till is thin or absent [*Alley et al.*, 1994b]. A hydrological potential map of Φ (Eq. 1) 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 limb joins the downglacier trunk (Fig. 1). *Alley et al.* [1994a] thus hypothesized that the limb remains active but aseismic because its sticky spots are water-lubricated, while the trunk nearly stopped and is seismically active on the sticky spots. The hydrologic potential Φ 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 overlying ice were ice-sheet ice (with the associated high surface slopes). Under those conditions, which *Alley et al.* [1994a] 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 Φ 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.

In 1994–1996 an experiment was conducted to test this hypothesis. Seismometers were deployed approximately every 90 km along the length of the ice stream 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 Km 10 is 10 km upstream of the grounding line, and so on). The rate of basal seismicity R (number of events per day) is low for the two sites on isB (UPB and DNB), and for sites Km 482 and Km 432 (in the catchment of isC, and in the uppermost part of isC, respectively). There is a marked increase in seismicity between Km 432 and Km 354 and seismicity remains high from Km 354 down to the array closest to the grounding line at Km 10 (Fig. 2). Flow velocities are low on the ice sheet (Km 482); the ice flows faster in the upper reaches of isC (Km 432 to Km 354: $30 < u < 60\,\mathrm{m \cdot a^{-1}}$) but nearly stagnates 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 has a relatively high velocity but also has high seismicity. This site is transitional between streaming and non-streaming ice and seems to exhibit some of the qualities of each.

Thermal processes have been suggested as controls on alternating fast and slow ice flow [*MacAyeal*, 1993b]. 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*, 1993b, a]. The thin ice of ice stream C, and the likelihood that it advected much inland cold ice near the bed, suggest the possibility of this process having acted on ice stream C (A Payne; B Kamb& HE Engelhardt, Chapman Conf., Orono, ME, 1998)

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 porosity 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.

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.

5 Conclusion

The rapid flow of ice streams depends on the 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. 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 as a result 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 called the limb, and existed under all of ice stream C as recently as 140–185 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 that $10\,\mathrm{m \cdot a^{-1}}$, and the ice stream is thickening at a rapid rate of up to $0.55\,\mathrm{m \cdot a^{-1}}$.

The mass balance of the combined ice stream B & C system (ice stream and catchment) appears to be thinning slightly 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 the other active ice streams, D, E, and F [*Shabtaie and Bentley*, 1987] is connected to general headward extension of the ice streams that resulted in the accidental triggering of the shutdown of C.

This shutdown of ice stream C is not an inherent feedback mechnism stabilizing the ice sheet, but a consequence of an accident of bed topography. 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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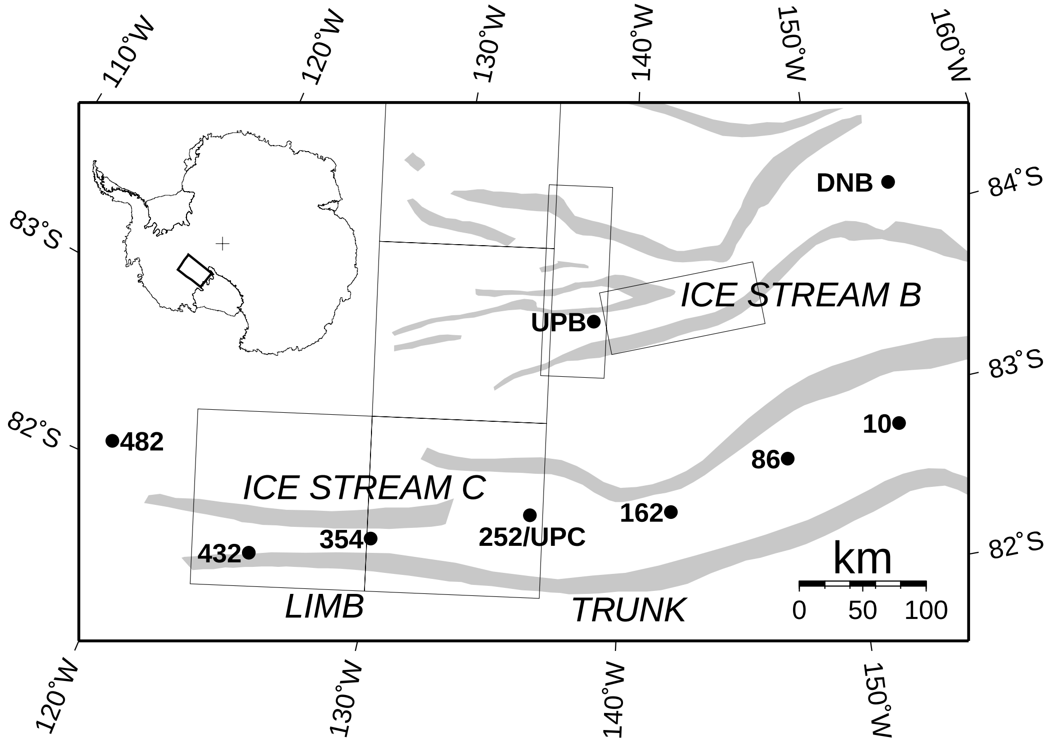
Figure 1: (a) Locations of the arrays of high-frequency seismometers on ice streams C and B. Flow is from the left to the Ross ice shelf on the right. The site name is the approximate distance in kilometers from the grounding line (which is from *Shabtaie et al.* [1987b]). Upstream B camp (UPB) and Downstream B camp (DNB) on ice stream B are also shown. Inset is a location map of Antarctica with the study region outlined. The ice-stream margins are from *Shabtaie and Bentley* [1987]. The outlined boxes are the region of radar coverage of *Retzlaff et al.* [1993]. (b) A zoomed view of the upstream regions of ice streams C and B. The basal hydrologic potential ϕ (Eq. 1) is contoured and our interpreted divides are shown in heavy curves. The dashed line is our extrapolation of a divide past the edge of the data.

Figure 2: Plot of ice-flow velocity u ($\mathrm{m \cdot a^{-1}}$) and seismicity R (day^{-1}) at the different array locations. Note that the values of $R = 0$ are plotted at $R = 0.1$ because of the log scale.

Figure 3: A satellite image of Siple Dome and the Duckfoot area. From *Jacobel et al.* [in review].


110°W
120°W
130°W
140°W
150°W
160°W
83°S
84°S
82°S
83°S
82°S
DNB
ICE STREAM B
UPB
10
482
86
ICE STREAM C
162
252/UPC
354
432
km
0 50 100
LIMB
TRUNK
120°W
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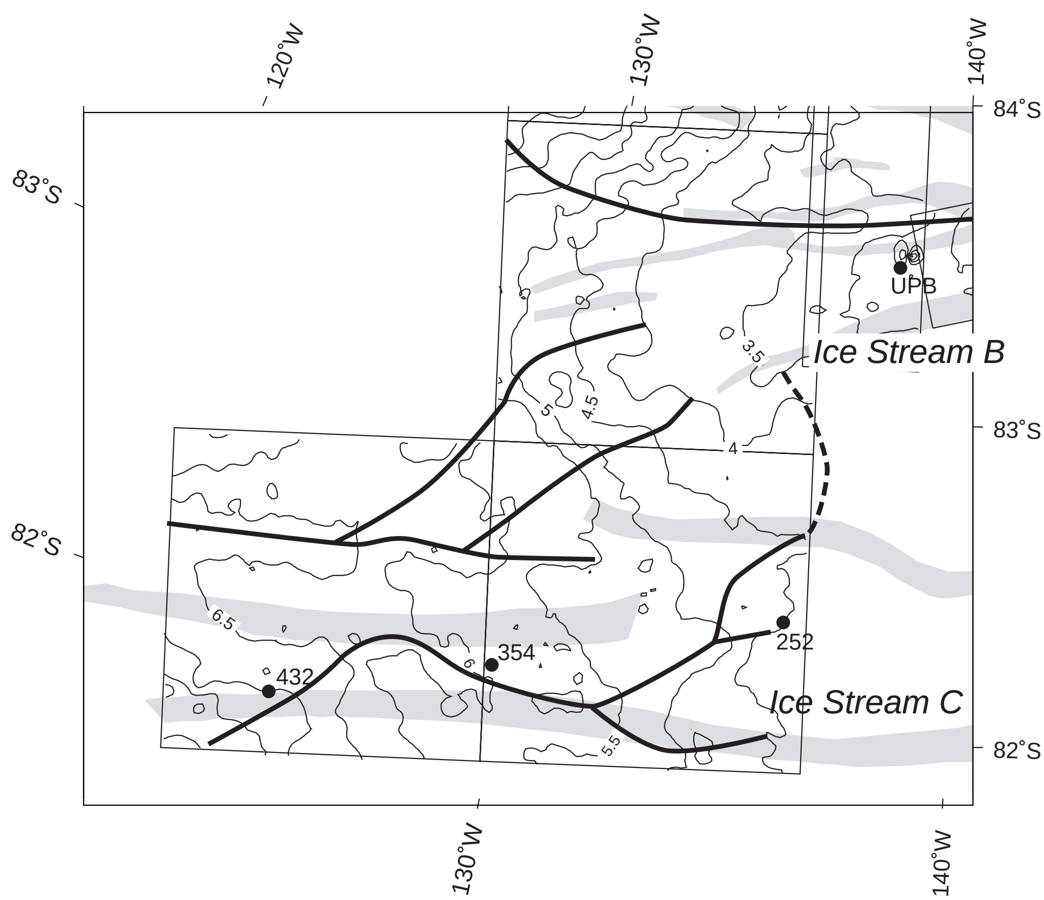
140°W
150°W


120°W
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Ice Stream B
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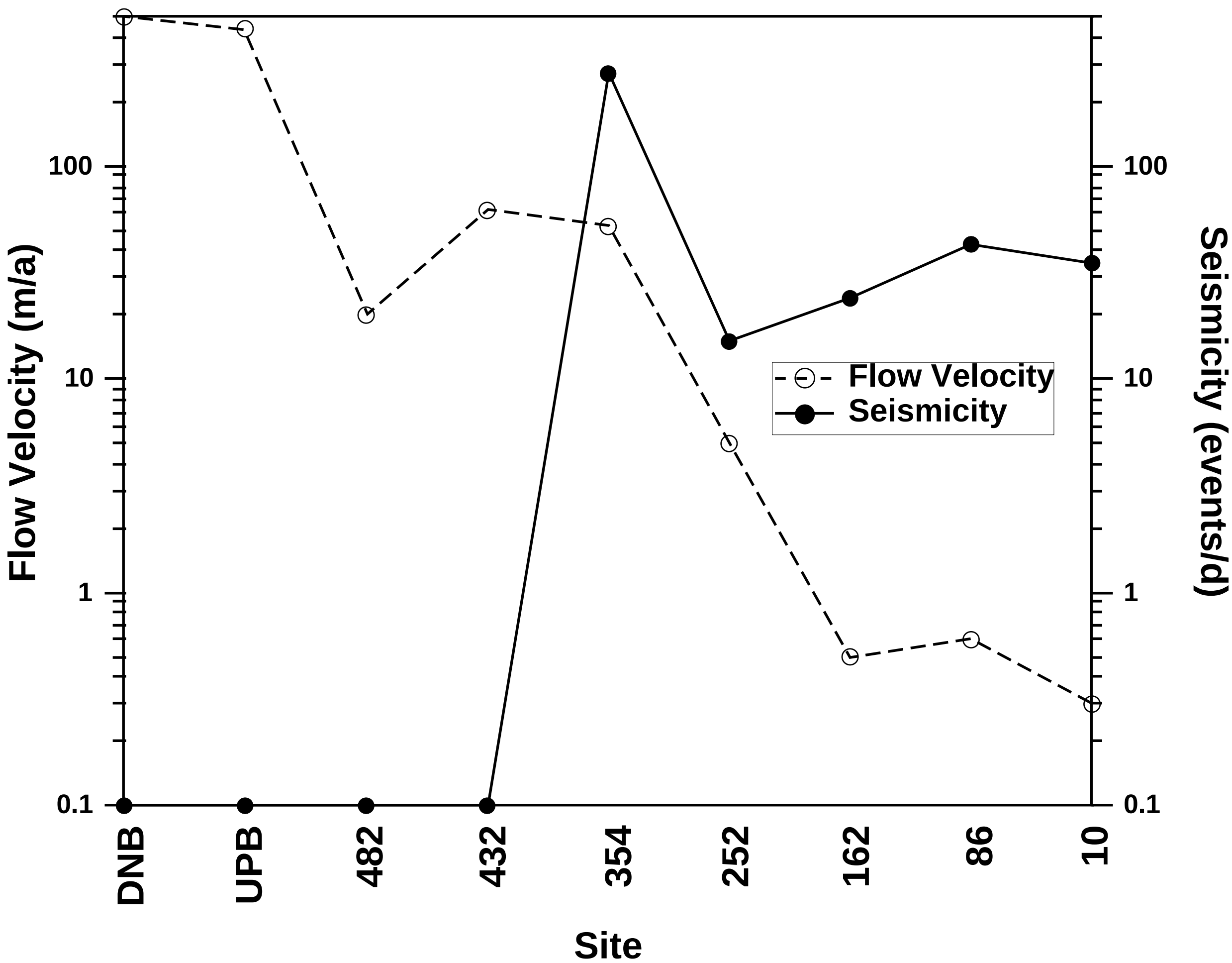
130°W
140°W


Flow Velocity (m/a)
Seismicity (events/d)
Flow Velocity
Seismicity
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0.1
DNB
UPB
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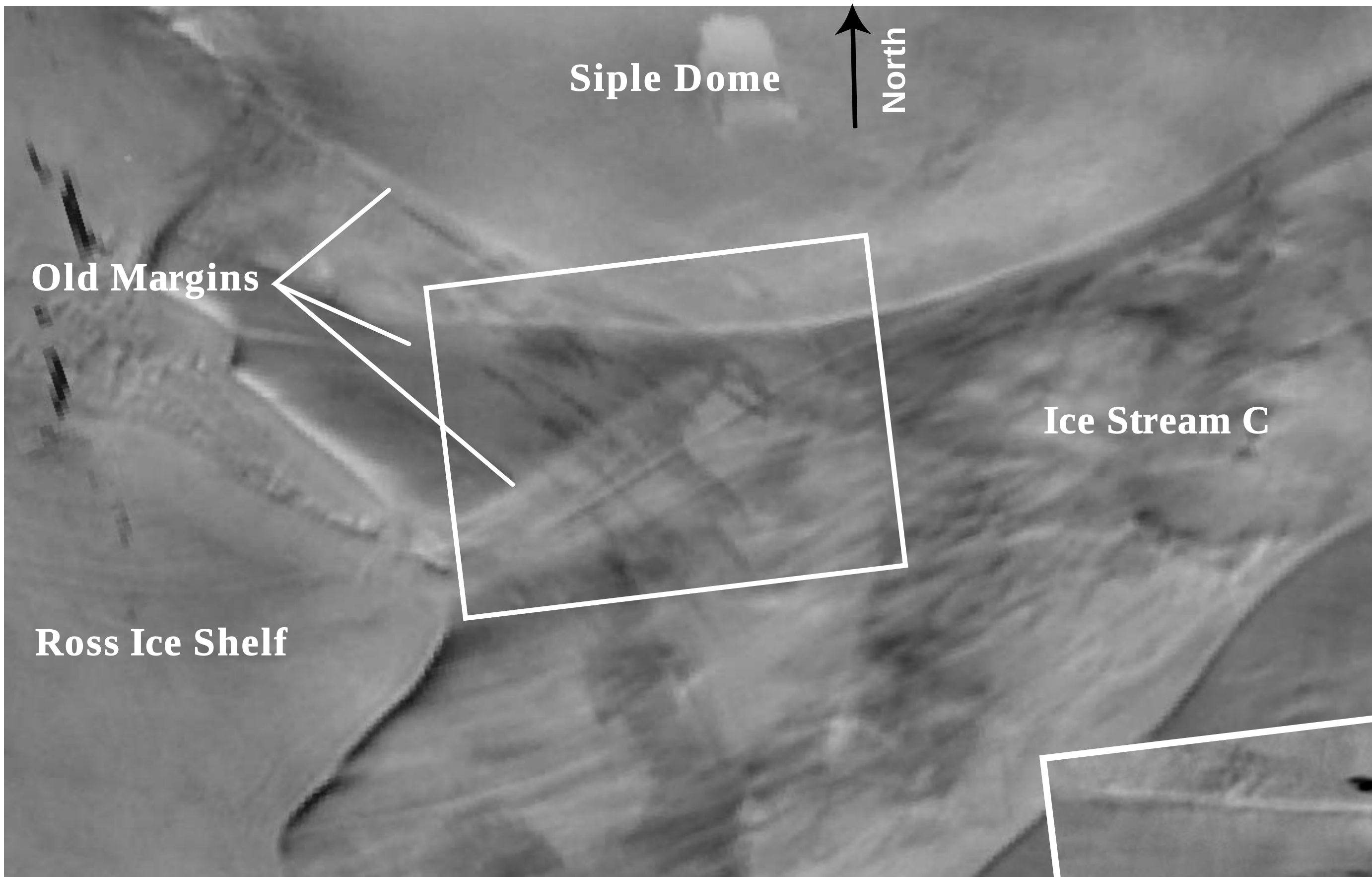
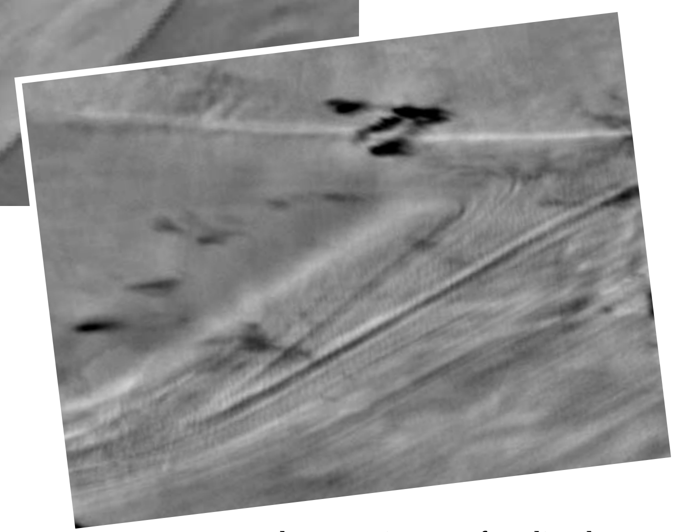
252
162
86
10
Site



AVHRR composite image (image size 220 km x 140 km)



Landsat TM image of outlined area