Ice Stream C, Antarctica, sticky spots detected by microearthquake monitoring

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ABSTRACT. Microearthquakes at the base of slow-moving Ice Stream C occur many times more frequently than at the base of fast-moving Ice Stream B. We suggest that the microearthquake source sites are so-called "sticky spots", defined as limited zones of stronger subglacial material interspersed within a weaker matrix. The fault-plane area of the microearthquakes ($O(10^2 \, m^2)$) is therefore a measure of the size of the sticky spots. The spatial density of the microearthquakes ($O(10 \, km^{-2})$) is a measure of the distribution of sticky spots.

The average stress drop associated with these microearthquakes is consistent with an ice-stream bed model of weak subglacial till interspersed with stronger zones that support much or all of the basal shear stress. We infer a weak inter-sticky-spot material by the large distances ($O(10^3 \, m)$), relative to fault radius, to which the microearthquake stress change is transmitted.

1. INTRODUCTION

Ice Stream C, West Antarctica, one of the Ross ice steams, has slowed to speeds of between 1 and 13 m a^{-1} (Whillans and others, 1987) within the last two centuries (Rose, 1979; Shabtaie and Bentley, 1987; Retzlaff and Bentley, 1993). Neighboring Ice Stream B flows rapidly (400–800 m a^{-1}) even though it is grossly similar to Ice Stream C in physical dimensions, accumulation and temperature (Shabtaie and others, 1987). The likely candidates for the restraining stress are the sides and the bed of the ice stream. In this paper we concentrate on the bed.

The sources of restraining stress at the bed are some unknown combination of the rheological properties of the till and the so-called "sticky spots" at the glacial/ subglacial interface (Bindschadler and Scambos, 1991; Alley, 1993; Whillans and Van der Veen, 1993). The sticky spots are regions of the bed where shear stress is concentrated and that, in sum, balance some or all of the driving stress. The deforming till beneath Ice Stream B (Blankenship and others, 1987; Engelhardt and others, 1990) permits fast flow but may provide little restraint to the driving stress of Ice Stream B (Alley and Whillans, 1991). Thus, the sticky spots may be important in basal drag.

We assume that conditions beneath Ice Stream C, when it flowed rapidly (>130 a ago), were similar to those currently beneath fast-flowing Ice Stream B. Thus, by extension, differences between Ice Streams B and C today are analogous to differences between Ice Stream C today and its condition two centuries ago. Identifying the critical parameter(s) that differentiate Ice Streams B and C could aid in developing theories for the shut-down of Ice Stream C. One of the most striking differences between the two ice streams is the rate, distribution and size of basal microearthquake activity.

The basal events are two orders-of-magnitude more numerous, 1–2 orders-of-magnitude larger moment, and more widely spatially distributed beneath Ice Stream C as compared to Ice Stream B (Anandakrishnan and Bentley, 1993). The events beneath both ice streams share the same depth-zone (the base of the ice) and the same focal mechanism (low-angle thrust faults in the direction of local driving stress), suggesting that the principal directions of stress at the bed are similar in both cases. We do not insist on this interpretation because slip on pre-existing faults is not necessarily determined by the stress directions. Nonetheless, we suggest that the difference in microearthquake rate and size is due to a difference in material properties (most notably friction) and the difference in spatial density is due to a difference in either friction or till rheology.

Alley and others (1994) proposed that an important parameter in controlling the velocity of ice streams is the state of lubrication of sticky spots. They hypothesized that basal water has been diverted away from Ice Stream C, resulting in a disruption of a Weertman-type water film. This water film (in their hypothesis) existed when Ice Stream C flowed fast but is currently either non-existent or discontinuous. In either case, the lubrication of sticky spots is reduced and Ice Stream C shuts down.

As we now demonstrate, the till rheology (as determined by porosity) is unlikely to have been much affected by the disruption of the water film. Thus, the sticky spots are as important to the dynamics of Ice Stream C flow as they are for Ice Stream B.

We now demonstrate that dewatering the till (changing the porosity from 40 to 30%) by conduction

through the till requires $O(10^6 \text{a})$. The volumetric flow rate u through a permeable medium is given by Darcy's law as $u = -K \text{d}H/\text{d}x$, where K is the hydraulic conductivity $(K = 10^{-6} \text{m s}^{-1})$ and the hydrologic potential gradient $\text{d}H/\text{d}x = 10^{-3}$. The flow rate per unit width is $q = u \times h$, where h is the thickness of the permeable till. For a 10 m thickness of till, $q = 10^{-8} \text{m}^2 \text{s}^{-1}$, and to move 10% of the water to the grounding line (10^5m) would require $t = (10^5 \text{m} \times 10 \text{m} \times 0.1)/q = 10^6 \text{a}$.

If, on the other hand, channels were to form at the base of the ice stream, the till would dewater upwards and that water would be quickly removed. However, after the top few centimeters of till had dewatered, the hydraulic conductivity of that zone would decrease by many orders to magnitude (Boulton and Hindmarsh, 1987) and slow the dewatering process for the deeper till. The cap over the till would tend to maintain the porosity of the till on timescales on the order of centuries. We argue that an unconsolidated till at the bed of Ice Stream C that permitted fast ice flow 130 a ago would likely still exist. Supporting these theoretical considerations is the discovery of a 20 km wide band of presumed dilatant till beneath Ice Stream C flanked by till of indeterminate dilatancy (Atre and Bentley, 1993). These results are from an analysis of the phase-reversals of reflections from the bed during an explosive-source seismic profile on Ice Stream C.

2. ANALYSIS

During the austral summer of 1987–88, a seismic array recorded 161 thrust-faulting events from the base of Ice Stream C (Anandakrishnan and Bentley, 1993). All the events shared the same fault-plane solution of slip at an angle of 30° to the long axis of the ice stream at a dip of a few degrees. All the events were therefore subject to the same stress field that is due almost exclusively to the surface and basal slopes of the ice stream.

The events were within 10 m of the bed of the ice stream, and Anandakrishnan and Bentley (1993) argued that they are most probably either in the subglacial region or at the interface, rather than within the ice. They argued that under the low strain rates present at the base of Ice Stream C, and the high temperature (close to or slightly below freezing), it is unlikely that ice would fracture (Hobbs, 1974). Rather, the basal shear stress is transmitted to the bed. Events on the interface between the ice and sediment would have moments determined by the average source parameters of the two media. We consider this situation most likely and use a moment determination that is intermediate between one calculated for ice and one calculated for the sediment.

The threshold-detection algorithm for the Ice Stream C events was used in two modes: single-station trigger and dual-station trigger. The single-station trigger would initiate recording of all seismograms in the array when the trigger station detected an event. To discriminate against small events, we occasionally used dual-station triggering where two widely separated stations must both detect an event before recording was initiated. The single-station-trigger events were on average smaller (lower seismic moment) than the dual-station-trigger events, but still had excellent signal-to-noise ratios (SNR > 18 dB).

2.1. Time-series analysis

We extract from this data set a list of event-occurence times t_i, inter-event times $\tau_i = t_i - t_{i-1}$, and the associated successive-event horizontal separation $d_i = |\mathbf{r}_i - \mathbf{r}_{i-1}|$ where $\mathbf{r}_i$ is the horizontal position of event i and $|\cdot|$ denotes the magnitude of the enclosed vector. We include event i if events t_i and t_{i-1} were not separated by any down-time of the recording system. This reduced the set to 53 events.

The counting statistic $n(T)$ is defined as the number of events in a time window of length T (Papoulis, 1984). The probability distribution function $P(n;T)$ is a Poisson distribution if all the events are independent. The horizontal separation d_i we defined for each pair of events t_i, t_{i-1} is now used to test whether the events are described by a Poisson process at all scales. That is, we now examine whether there is some critical event separation D such that the sub-set described by $d_i \leq D$ is *not* distributed as a Poisson, but events described by $d_i > D$ *are* Poisson. Certainly $D < \infty$ must exist because events at some (possibly great) distance must be independent. If $D = 0$, then the microearthquake events are independent one from another at all distances. We call the distance D the "radius-of-influence" (ROI) of a microearthquake.

We test whether a distribution is Poisson by a χ^2 test and reject the hypothesis of Poisson if the goodness-of-fit parameter q is small. The goodness-of-fit parameter $q = Q(\chi^2 || \nu)$ is the probability that the observed chi-square will exceed the value χ^2 by chance even though the model is correct; the function Q is the χ^2 distribution function (Walpole and Myers, 1975). Because of the small sample size, we reject the Poisson hypothesis if q is $O(10^{-10})$ but accept the hypothesis for q that is $O(10^{-5})$. This is not entirely arbitrary; examination of Figure 1 demonstrates that q tends to fall into two regions: $q \gtrsim 10^{-5}$ and $q \ll 10^{-10}$.

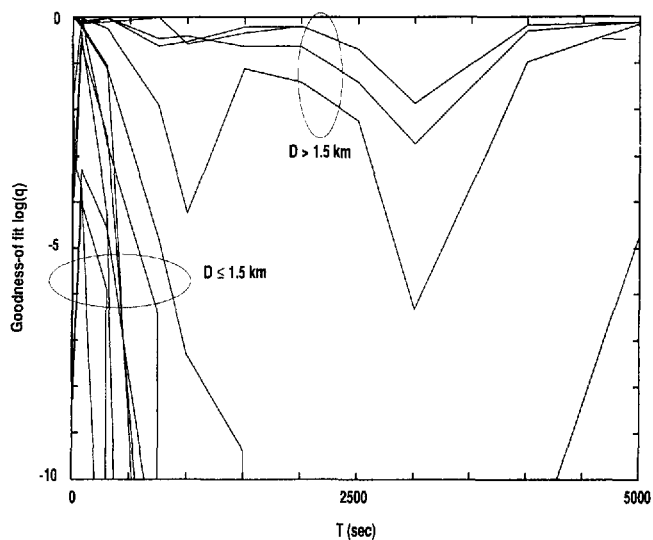The counting distribution $n(T)$ depends on the

Fig. 1. Plot showing log of the χ^2 goodness-of-fit parameter q versus the counting time T. The different curves are for different values of D, the critical distance between successive events.

counting length T. For T short compared to mean inter-event time τ, any distribution will be well described by a Poisson (Papoulis, 1984). Thus, we test for Poisson distributions by varying T from $T \ll \tau$ to $T \gg \tau$. If we cannot reject the Poisson hypothesis at all T, then we must accept it.

Figure 1 is a plot of q vs counting time T. The range of T is from 1 s to 5000 s with $\bar{\tau} \approx 800$ s. The different lines are for the sub-set $d_i < D$, with D ranging from 100 m to 2500 m. For $D \leq 1500$ m, we reject the hypothesis of Poisson distribution because $q \ll 10^{-5}$ at some T (generally at $T \gtrsim \tau$, q drops sharply from $O(10^{-5})$ to $O(10^{-20})$ or less). Note that, in accordance with theory, $q = 1$, indicating a good fit with the Poisson distribution at $T = 1$ s (a time much shorter than the mean inter-event time $\bar{\tau}$). For $D > 1500$ m, the hypothesis of Poisson distribution cannot be rejected. The threshold level of $q = 10^{-5}$ is somewhat arbitrary, but it is clear that for the three curves marked $D > 1500$, the hypothesis of Poisson is more plausible than for the other curves, which are marked $D \leq 1500$ m. Though the number of data points for the large-separation curves ($D > 1500$ m) is small ($n = 15$), the goodness-of-fit parameter q is large ($\gtrsim 10^{-2}$) at most T and never drops to 10^{-20} as do the curves marked $D \leq 1500$ m.

2.2. Spectral characteristics

Using the spectrum of the compressional-wave arrival, Anandakrishnan and Bentley (1993) calculated the seismic moment of the events using the methods of Aki and Richards (1980) and Brune (1970). The seismic moment calculation is a function of source density and shear-wave velocity. Assuming source parameters intermediate between ice and sediment, we calculate that the seismic moments M_0 for the 161 events beneath Ice Stream C averaged $\bar{M}_0 = 10^7$ N m.

Frequency analysis of the seismic wave forms using the method of Brune (1970) and Madariaga (1983) can be used to calculate the source radius. In one generally used method (Aki and Richards, 1980), the spectrum of the displacement wave form is fitted to a function of the form $u_0/(1 + (\omega_0/\omega)^2)$ where u_0 is related to the moment (and is a robust measure thereof) and ω_0 is related to the fault-plane area S. The angular frequencies ω and ω_0 are defined as $\omega \equiv 2\pi/f$ and $\omega_0 \equiv 2\pi/f_0$, where f and f_0 are spectral frequencies in Hz. Calculations of fault-plane area are highly model-dependent and therefore fault area S is not as well constrained as moment M_0, but is good to an order-of-magnitude. We estimate the fault area as $S \sim 10^2$ m^2.

Finally, using the moment and fault-plane area, we calculate the stress drop associated with the events by $\Delta\sigma \approx M_0/aL^3 \approx 10^5$ Pa where $L \sim \sqrt{S}$ is the characteristic dimension (radius or length) of the fault and $a = 0.1$ is a fudge factor for fault shape and could go to 0.3 (Kostrov and Das, 1989). Again, because of the uncertainty in L, the stress-drop $\Delta\sigma$ is an order-of-magnitude approximation.

Using simple Darcy's law arguments, we demonstrated that the till is likely to retain a high porosity regardless of the state of the basal water system. The results from the microearthquake time-series analysis suggest the same conclusion. The ROI for events at the bed of Ice Stream C is 1.5 km. A qualitative interpretation of this result is that the intervening material has low strength. Work in progress is aimed at quantifying a relation between ROI and sediment rigidity.

We now estimate whether the driving stress of the ice streams is balanced entirely by the stress drops of the microearthquakes. We assume that the bed alone supports the driving stress of the ice stream and ignore the role of the sides of the ice stream.

The shear stress at the base of the ice is given by $\sigma_i = \rho g h \alpha \approx 0.01$ MPa (Paterson, 1981) where $\rho = 10^3$ kg m^{-3} is the density of ice, $g = 10$ m s^{-2} is the acceleration due to gravity, $h = 10^3$ m is the ice thickness and $\alpha = 10^{-3}$ is the surface slope. The mean separation of events is approximately 400 m. If we assume that the inter-sticky-spot material is weak, then the stress of a $(400 \text{ m})^2$ area is concentrated on one sticky spot (on average). The force on each such sticky spot is $F = (400 \text{ m})^2 \times 0.01$ MPa $= 1.6 \times 10^9$ N. The area of each of the sticky spots is $S \sim 10^2$ m and thus the shear stress on each sticky spot is $\sigma = F/S \sim 10$ MPa. The seismically measured stress drop is 0.1 MPa.

Theories of partial stress drop hypothesize that an earthquake only releases a fraction of the total applied stress on the fault. Estimates of the fraction vary from 1 to 10% (Brune, 1970). Thus, if partial stress drop is 1%, then the driving force of the ice stream is entirely balanced by the sticky spots; if it is 10% then only 10% of the driving force is balanced by sticky spots.

3. CONCLUSION

Microearthquakes at the bed of Ice Stream C occur at discrete zones of high friction that are separated by low-strength till. These sticky spots balance most, if not all, of the driving stress of the ice stream and are strong candidates for the mechanism for restraining ice flow. We have demonstrated that, even though the fault radius of an individual microearthquake event is $O(10 \text{ m})$, the events can modify the stress at locations up to 1.5 km away. This argues for a weak intervening material.

For sticky spots with a mean separation of 400 m, a mean radius of 10 m and with a weak intervening till, the sticky spots support 10–100% of driving stress for ice flow. Thus, we suggest a model of the base of the ice streams that consists of a weak, unconsolidated till interspersed with discrete zones of higher strength that are on the order of 10^2 m^2 in area. These zones have a spatial density on the order of 10 km^{-2}. The frictional character of these sticky spots is a possible candidate for the controlling parameter for ice-stream flow speed.

The suggested diversion of Ice Stream C basal water (Alley and others, 1994) and the consequent slowing of Ice Stream C is likely to have preserved the porosity of the till and the size and density of the sticky spots. The friction on the sticky spots is the only parameter likely to have changed. We suggest that sticky spots exist beneath Ice Stream B but only rarely do they manifest themselves seismically. The microearthquakes emanating from the base of Ice Stream C are the first direct evidence of the difference in frictional character of sticky spots between the fast and slow ice streams.

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REFERENCES

Aki, K. and P.G. Richards. 1980. *Quantitive seismology. Theory and methods.* New York, W.H. Freeman and Co.

Alley, R.B. 1993. In search of ice-stream sticky spots. *J. Glaciol.*, **39**(133), 447 454.

Alley, R.B. and I.M. Whillans. 1991. Changes in the West Antarctic ice sheet. *Science*, **254**(5034), 959-963.

Alley, R.B., S. Anandakrishnan, C.R. Bentley and N. Lord. 1994. A water-piracy hypothesis for the stagnation of Ice Stream C, Antarctica. *Ann. Glaciol.*, **20** (see paper in this volume).

Anandakrishnan, S. and C.R. Bentley. 1993. Micro-earthquakes beneath Ice Streams B and C, West Antarctica: observations and implications. *J. Glaciol.*, **39**(133), 455-462.

Atre, S.R. and C.R. Bentley. 1993. Laterally varying basal conditions under Ice Streams B and C, West Antarctica. *J. Glaciol.*, **39**(133), 507-514.

Bindschadler, R.A. and T.A. Scambos. 1991. Satellite-image-derived velocity field of an Antarctic ice stream. *Science*, **252**(5003), 242-246.

Blankenship, D.D., C.R. Bentley, S.T. Rooney and R.B. Alley. 1987. Till beneath Ice Stream B. I. Properties derived from seismic travel times. *J. Geophys. Res.*, **92**(B9), 8903-8911.

Boulton, G.S. and R.C.A. Hindmarsh. 1987. Sediment deformation beneath glaciers: rheology and geological consequences. *J. Geophys. Res.*, **92**(B9), 9059-9082.

Brune, J.N. 1970. Tectonic stress and the spectra of seismic shear waves from earthquakes. *J. Geophys. Res.*, **75**(26), 4997-5009.

Engelhardt, H., N. Humphrey, B. Kamb and M. Fahnestock. 1990. Physical conditions at the base of a fast moving Antarctic ice stream. *Science*, **248**(4951), 57 59.

Hobbs, P.V. 1974. *Ice physics.* Oxford, Clarendon Press.

Kostrov, B.V. and S. Das. 1989. *Principles of earthquake source mechanics.* Cambridge, Cambridge University Press.

Madariaga, R. 1983. Earthquake source theory: a review. *In* Kanamori, H., *ed. Earthquakes: observation, theory and interpretation.* New York, International School of Physics "Enrico Fermi", North-Holland Publishing, 1-41.

Papoulis, A. 1984. *Probability, random variables and stochastic processes. Second edition.* New York, McGraw-Hill.

Paterson, W.S.B. 1981. *The physics of glaciers. Second edition.* Oxford, Pergamon Press.

Retzlaff, R. and C.R. Bentley. 1993. Timing of stagnation of Ice Stream C, West Antarctica, from short-pulse radar studies of buried surface crevasses. *J. Glaciol.*, **39**(133), 553-561.

Rose, K. 1979. Characteristics of ice flow in Marie Byrd Land, Antarctica. *J. Glaciol.*, **24**(90), 63-75.

Shabtaie, S. and C.R. Bentley. 1987. West Antarctic ice streams draining into the Ross Ice Shelf: configuration and mass balance. *J. Geophys. Res.*, **92**(B2), 1311-1336.

Shabtaie, S., I.M. Whillans and C.R. Bentley. 1987. The morphology of Ice Streams A, B, and C, West Antarctica, and their environs. *J. Geophys. Res.*, **92**(B9), 8865-8883.

Walpole, R.E. and R.H. Myers. 1975. *Probability and statistics for engineers and scientists. Second edition.* New York, Macmillan.

Whillans, I.M. and C.J. van der Veen. 1993. New and improved determinations of velocity of Ice Streams B and C, West Antarctica. *J. Glaciol.*, **39**(133), 483 490.

Whillans, I.M., J. Bolzan and S. Shabtaie. 1987. Velocity of Ice Streams B and C, Antarctica. *J. Geophys. Res.*, **92**(B9), 8895-8902.

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