Microearthquakes and subglacial conditions

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[1] Ten passive seismic recording stations were deployed in a 9 km by 6 km array on Rutford Ice Stream, West Antarctica, to detect microearthquakes from the ice stream bed. The array covered an area where a varying distribution of basal sedimentary conditions had been identified from seismic reflection surveys and interpreted as areas of both basal sliding and subglacial sediment deformation. The passive seismic experiment was designed to give an independent confirmation of this interpretation, as basal sliding is believed to be associated with the release of more acoustic energy than a deforming bed. The results show that this is the case, basal sliding producing six times more events. This confirms that the spatial pattern of basal conditions does indeed reflect the distribution of different ice flow mechanisms. It also shows that microearthquake monitoring can be a valuable new technique for mapping ice stream basal conditions over wide areas.
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1. Introduction

[2] Natural seismic emissions from glaciated regions have been recorded many times. For example, seismic events associated with both the internal stress regime and with ice motion over the bed have been reported from a number of glaciers [e.g., *Deichmann et al.*, 2000; *Neave and Savage*, 1970; *VanWormer and Berg*, 1973; *Weaver and Malone*, 1976]; emissions associated with a glacier surge have been observed [*Stuart et al.*, 2005]; and the ability of glaciers to generate magnitude 5 earthquakes has been reported [*Ekström et al.*, 2003]. In West Antarctica, *Blankenship et al.* [1987] first recorded the natural seismicity from an ice stream bed, referring to the events as microearthquakes. Other studies have interpreted their source mechanisms and characteristics, showed correlations with ice flow and with ocean tides, compared different ice streams, and interpreted mechanisms of flow control and evolution [*Anandakrishnan and Bentley*, 1993; *Anandakrishnan and Alley*, 1994, 1997a, 1997b; *Anandakrishnan et al.*, 2001]. This paper uses new microearthquake data from Rutford Ice Stream, West Antarctica to infer subglacial conditions, with control from previous seismic reflection results.

[3] The flow of ice streams in West Antarctica is of particular interest for the significant role they are believed to play in the overall dynamics of the ice sheet and any changes it may undergo in future. Rutford Ice Stream (Figure 1) has been the site of considerable glaciological research over the past 30 years [*Doake et al.*, 2001], including the use of seismic reflection techniques to investigate the ice-stream bed [e.g., *Smith*, 1997a, 1997b, 1997c; *Vaughan et al.*, 2003]. Those studies determined the acoustic impedance of the water-saturated sediment bed beneath the ice and hence, an estimate of its porosity. A range of porosities was found with a largely bimodal distribution [see *Smith*, 1997a, Figure 6]. Some porosity values matched what would be expected from dilated sediment ($\sim$0.4), whilst others were more typical of un-dilated sediment ($\sim$0.3 or less). The ice stream bed comprises adjacent patches of dilated and un-dilated sediment, hundreds of meters to kilometers wide, separated by very sharp boundaries [*Smith*, 1997a]. In some places this pattern was mapped and the different patches appear to be elongated in the ice flow direction. Repeated surveys have shown the boundaries moving (but still remaining sharp), rapid erosion ($1\ \mathrm{m\ a^{-1}}$) of un-dilated sediment and moulding of dilated sediment into new bed forms [*Smith et al.*, 2006]. Following *Atre and Bentley* [1993], a model was proposed in which the dilated sediment was interpreted as deforming pervasively with the motion of the overriding ice, whereas in areas underlain by un-dilated sediment, the ice flows mainly by sliding over the bed, with perhaps some shear deformation within the upper sediment layers [*Smith*, 1997a].

[4] Further surveys, including ones on other ice streams and glaciers, have shown the model to be robust and consistent [e.g., *Smith et al.*, 2002; *Vaughan et al.*, 2003; *King et al.*, 2004]. It was also used to show how, when averaged over wide areas, the basal conditions control the restraint exerted by the bed on the ice flow [*Vaughan et al.*, 2003]. However, independent confirmation of this distinction between deforming bed and basal sliding has been lacking. A passive seismic network was deployed on Rutford Ice Stream in December 1997 with the aim of providing this confirmation. Both theoretical and observational considerations suggest that a reduction in sediment porosity from 0.4 to 0.3 should result in an increase in seismicity, as homogeneous sediment deformation changes to localized shearing [for a summary, see *Anandakrishnan and Bentley*, 1993]. The experiment therefore aimed to determine whether or not the low-porosity bed areas, interpreted as basal sliding from the seismic reflection data, showed a significantly higher level of seismicity than the high-porosity ones, believed to be associated with a deforming sediment bed.

2. Data Acquisition

[5] A network of ten passive seismic stations was deployed approximately 40 km upstream of the grounding line of Rutford Ice Stream (Figure 1). Each station comprised a 3-component geophone (natural frequency 20 Hz)

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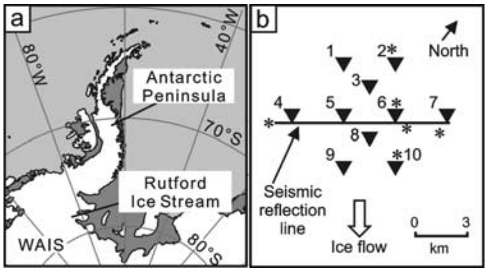

Figure 1. Location maps. (a) Rutford Ice Stream, West Antarctica. WAIS: West Antarctic Ice Sheet. (b) Microearthquake recording array geometry (plan view). Triangles show the recording station locations. Black line passing through stations 4–7 shows the location of the seismic reflection line of *Smith* [1997a]. Mean ice flow velocity across the array is 378 m a^{-1} (stars show the location of the velocity measurements). The ice stream shear margins begin approximately 5 km beyond stations 4 and 7.

and Reftek DAS data logger. One geophone component was orientated vertically, one was parallel to ice flow and the third was perpendicular to ice flow. Recording was initiated by the detection of an event on the vertical channel and the data were sampled at 500 Hz. The array operated continuously for a period of ~11 days. Individual stations in the array were positioned as indicated in Figure 1. Four stations (4–7) were located along a line where the distribution of subglacial conditions had previously been interpreted using seismic reflection techniques [*Smith*, 1997a]. Part of that earlier seismic reflection line (a 3.6 km section, including stations 5 and 6) was repeated during the passive seismic experiment. This showed there had been no changes in the basal conditions during the intervening period. The other passive seismic stations were distributed upstream and downstream of the seismic reflection line. Station spacing on the main array was 3 km, which is approximately 1.5 times the ice thickness. Stations 3 and 8 were located at slightly closer spacing, near the center of the main array. Ice flow velocity across the array varied by <4% so levels of seismicity at the different sites can be compared directly, without the need to consider any flow differences.

3. Microearthquake Event Identification and Source Location

[6] The events of particular interest were those originating at, or close to the ice-bed interface. Those from all other sources (e.g., surface events and camp noise) were discarded [*Blankenship et al.*, 1987; *Anandakrishnan and Alley*, 1997b]. A typical bed event is shown in Figure 2. Bed events recorded at all the stations were short duration and low amplitude, though still well above the background noise level. Spectra of the vertical components show a broad range of frequencies from 50 Hz to 250 Hz (the Nyquist frequency), with a peak at around 200 Hz. The vertical component arrival was followed, a short time later, by distinct arrivals on the horizontal channels. Approximate P- and S-wave velocities in ice are well known (~3600–3900 m s^{-1} and ~1700–1950 m s^{-1}, respectively) [e.g., *Roethlisberger*, 1972]; and ice thickness in much of the area is known from seismic and radar surveys. For each event, the difference in travel-time between arrivals on the different channels is consistent with P and S waves originating from a single source at the ice stream bed. P- and S-wave travel-time differences range between 650 and 750 ms. The minimum value is consistent with a source almost directly beneath the recording station; the maximum value indicates a source with an epicentral distance from the station of ~1 km. A similar pattern of P and S wave arrivals is also seen on other ice streams [e.g., *Blankenship et al.*, 1987; *Anandakrishnan and Bentley*, 1993; *Anandakrishnan and Alley*, 1997a].

[7] No bed events could be identified conclusively on more than one recording station. This observation, combined with the travel-time and waveform characteristics, has significant implications. Events from the ice stream bed are low energy, otherwise they should have been detected at more than one station. They are strong enough to trigger the nearest station, but not those further away. At least in this region of the ice stream, each station was sensitive only to events occurring within an epicentral radius of ~1 km. Without making assumptions about the source mechanism, it is not possible to determine the direction of each epicenter from the station position, thus precluding an accurate location of the source. All that can be determined is the position of a circle on the bed on which the source lies. This is the maximum area of the bed to which a station was sensitive and each station was effectively independent of the others in the array. Depending on the source mechanism and the potential for focusing of upward traveling waves, the true area of sensitivity may be more restricted than this [*Anandakrishnan and Bentley*, 1993], but we cannot conclude that for certain from these data alone. Closer station spacing and continuous recording would probably have been required to determine source locations accurately.

4. Event Repeatability and Periodicity

[8] Basal seismicity at all sites showed discrete, periods of activity lasting up to 12 hours. Each period of activity began and ended abruptly and was followed by a quiet period of between a few hours and many days, during which events were detected rarely, if at all. Mean basal seismicity varied across the array (Figure 3). At the sites with higher mean values, the active periods occurred more often, generally lasted longer, and within them events were more frequent, than at other sites. In the longer, active periods at

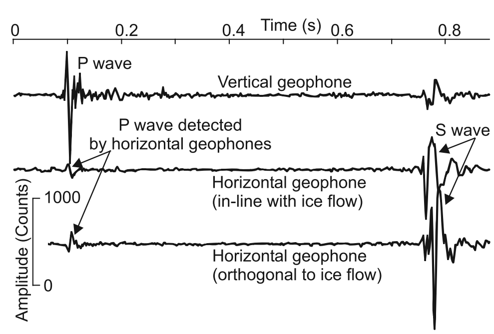

Figure 2. Example of a typical microearthquake from the bed of Rutford Ice Stream. Vertical scale is the same for all three traces (units are counts from the data logger).

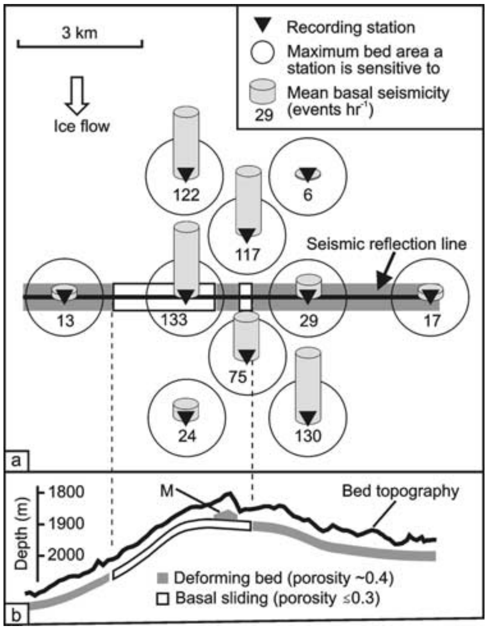

Figure 3. Basal seismicity from the microearthquake array compared with basal conditions from seismic reflection data. (a) Mean basal seismicity at each recording station. (b) Distribution of deforming bed and basal sliding from *Smith* [1997a]. This is repeated in Figure 3a, where it coincides with the bed directly beneath stations 4–7. M is a large mound of deforming sediment on top of non-deforming bed. Bed topography scale is depth below sea level.

all the sites, three characteristics of the seismic events are significant. First, many consecutive events had virtually identical waveforms, even for hundreds of events spread over many hours; second, the time delay between the P- and S-wave arrivals for these events was identical (Figure 4a); and third, the time between each consecutive event was very similar (Figure 4b). At the different sites, as well as at different times within a site, the wavelet shapes and these time separations were often different to those in Figure 4, but these three characteristics were the same within a particular group of events. Henceforth, these groups of events will be called swarms. Figure 4a also shows the discrimination between two different swarms detected at Station 1. The events with a P-S separation of 672 ms had a waveform slightly, but consistently, different to those in the 728 ms swarm. These are interpreted as different swarms, the shorter P-S separation indicating a source with an epicenter closer to the recording station.

[9] Basal events on ice stream C (West Antarctica) have been interpreted as resulting from stick-slip motion along low-angle faults at the ice-bed interface or within the bed itself [*Anandakrishnan and Bentley*, 1993]. The spectra of the vertical components from Rutford Ice Stream show a very rapid fall-off just before the Nyquist frequency (250 Hz), indicating that the data are under-sampled and that event waveforms cannot be recreated reliably. This precludes calculations of the seismic moment or a similar determination of the source mechanism from these data. However the similarities in the waveforms suggest that the same mechanism could be involved. The arrival separation times in Figure 4a are accurate to ±2 ms which suggests that sources more than 12 m apart could be distinguished and identified as coming from different locations on the bed. However, no events in this swarm show P-S separation times outside these error limits (Figure 4a), which is consistent with *Anandakrishnan and Bentley*'s [1993] interpreted fault plane dimensions of around 10 m.

[10] The repeatability and periodicity characteristics suggest that each swarm involves a sequence of small, identical source events, occurring periodically at the same place. This suggests accumulated stress may be released slowly, at one location, over a period of many hours. Only a small number of swarms occurred at each site, implying that activity at only a few discrete locations contribute most of the recorded seismicity.

5. Microearthquakes and Subglacial Conditions

[11] Mean basal seismicity determined at each station is given in Figure 3, which also shows the interpretation of the seismic reflection line [*Smith*, 1997a] passing through stations 4–7 of the array. These four sites show a good correlation between basal seismicity and the different con-

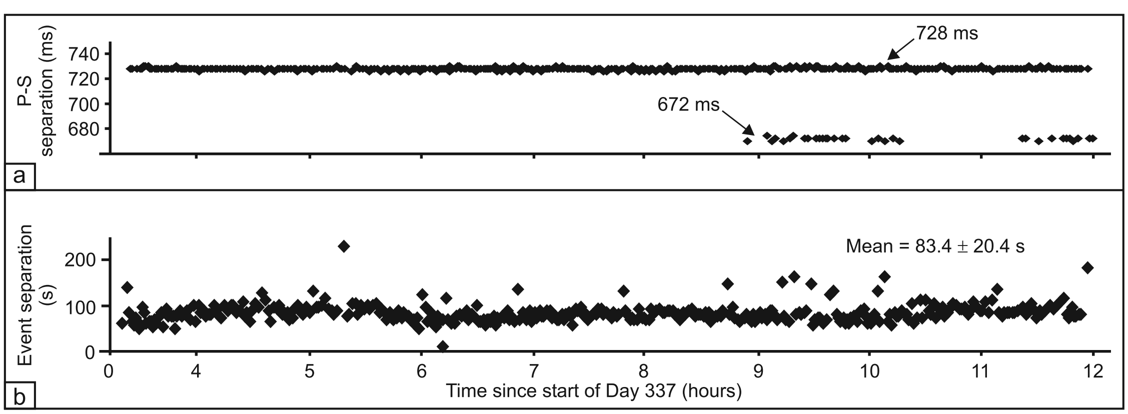

Figure 4. Repeatability and periodicity of individual event swarms. Example of 12 hours of data from Station 1. Errors in time values are ±2 ms. (a) Time difference between arrival of the P and S waves for individual events. Two separate swarms are indicated, with time differences of 728 ms (373 events) and 672 ms (33 events). Within each swarm, all events have virtually identical waveform characteristics. (b) Time difference between consecutive events (P wave arrivals) for the 728 ms swarm.

ditions interpreted at the bed. Basal sliding is associated with a level of seismicity approximately six times greater than a deforming sediment bed (i.e., comparing station 5 with the mean of stations 4, 6, and 7). This confirms the model used to interpret the seismic reflection data and suggests passive seismic arrays can be used to map the distribution of these different basal conditions and flow mechanisms. However, this result cannot necessarily be taken as evidence that basal sliding is also a mechanism of greater restraint to ice flow. Strictly, it shows only that basal sliding is associated with a higher degree of elastic energy release than deforming bed. It is possible that viscous deformation of the basal sediment will also be a mechanism of significant resistance to ice flow, but one without the release of elastic energy and hence without a high level of observed seismicity. Although *Vaughan et al.* [2003] showed that, when averaged over a wide area (3–5 ice thicknesses), basal conditions do influence the ice flow, with basal sliding areas exerting greater restraint, more detailed experiments would be required to show if this remains true at the scales associated with the passive seismic observations too (0.5–1 ice thickness). It is also worth noting that we still cannot determine whether the motion in basal sliding areas is concentrated actually at the ice-bed interface, or else on shear planes within the bed.

6. Conclusions and Discussion

[12] At this location on Rutford Ice Stream, ice flow by basal sliding is associated with a level of seismicity six times greater than the deforming bed mechanism. Monitoring basal seismicity is a potentially useful technique to distinguish between these different basal conditions and associated ice flow mechanisms. This is independent confirmation that high porosity, dilated sediment is undergoing pervasive deformation (the deforming bed ice flow mechanism), whereas un-dilated, lower porosity sediment (basal sliding), is not. It also further supports identification of the close juxtaposition and discrete boundaries of the different ice flow mechanisms interpreted from seismic reflection data. This result, by improving the understanding of the subglacial environment, has implications for modeling both steady-state behavior and temporal evolution of ice streams and glaciers.

[13] The correlation between basal conditions and seismicity where the two techniques overlap (stations 4–7) is very good. This suggests that basal sliding also dominates beneath stations 1, 3, 8, and 10, whereas stations 2 and 9 have deforming beds. Monitoring microearthquakes may thus be a powerful technique for mapping the distribution of basal conditions (deforming bed *vs*. basal sliding) over a wide area. Once the local levels of seismicity have been compared with known basal conditions (from seismic reflection surveys), this calibration can be extended over a much greater area using a network of passive seismic stations, which are far less labor-intensive. Alternatively, passive seismic stations may be used as a reconnaissance tool to identify critical locations for seismic reflection lines.

[14] Previous ice stream microearthquake experiments on the Siple Coast of West Antarctica [e.g., *Blankenship et al.*, 1987; *Anandakrishnan and Bentley*, 1993] developed many of the concepts and the underlying theoretical basis for the results presented here, and there are broad similarities between the two regions. Fast-flowing, soft-bedded Ice Stream B, which may have a considerable deforming bed component, is very quiet (0–0.1 events hr^{-1}) [*Anandakrishnan and Alley*, 1997a]. Beneath Ice Stream C, which flows much slower and beneath which little, or no deforming bed is interpreted, events are two orders of magnitude more frequent (up to 13 events hr^{-1}) [*Anandakrishnan and Alley*, 1997a], though this is still comparable only with the quieter sites on Rutford Ice Stream. However, any more-detailed comparisons are probably unjustified, as differences exist in both the glaciological regimes (e.g., ice thickness) and the data acquisition (e.g., sampling and triggering strategies). More importantly, the Rutford Ice Stream array specifically targeted known, detailed variations in basal conditions, whereas those on ice streams B and C concentrated on regional distributions as well as on event mechanisms. In this way, the focus on specific basal conditions on Rutford Ice Stream demonstrates a new application of the microearthquake monitoring technique in cryospheric research.

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