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Supplemental

Previous Results

The WDB is located in eastern Ross Sea and includes the continental shelf area north of the Ross Ice Shelf calving front that is bound by the Hayes and Houtz Bank. This large basin is approximately 250 km long in a north-south direction and approximately 100 km wide. The average depth of the basin is 600 m but there is a broad bathymetric saddle that rises to 450 m at a distance of approximately 70 km from the continental shelf edge. Mosola and Anderson (2006) used a dip-oriented seismic line and multibeam bathymetric data to demonstrate that grounded ice had advanced to the shelf edge during the LGM. Following the advance to the continental shelf edge, the retreat of grounded ice was interrupted and a large-volume GZW that corresponds to the bathymetric saddle was deposited. A high-resolution multibeam bathymetric survey and additional seismic data collected during expedition *NBP1502B* showed that the retreat of grounded ice was far more complex than could be resolved with the previously available data (Bart et al., 2017a). Mapping of the new and archived multibeam bathymetric data combined with correlations from relatively dense grid of seismic data revealed that the bathymetric saddle actually represents a stack of at least seven overlapping GZWs that were deposited after grounded ice began to retreat from the continental shelf edge (Bart et al., 2017a,b).

The seaward edges for four of the WDB GZWs are well exposed on the outer-most continental shelf and hence the former locations of the grounding zone are well resolved (Bart et al. 2017b). The imaging of seafloor morphology from the multibeam data allowed the exposed GZWs to be specifically cored on the GZW foreset surface. These core locations were obviously former sites of proximal grounding zone sedimentation. Indeed, the cores acquired on the foresets of the GZW1, -2 and -3 (the first three GZWs deposited after the ice retreated from the continental shelf edge) all bottom into homogenized glacial till (McGlannan et al., 2017). These core sites show an upcore sedimentary facies transition to weakly laminated sediments with abundant clasts. These sedimentary facies contain abundant benthic and planktonic foraminifera and indicate the accumulation of sediment in a sub-ice-shelf environment (Bart et al., 2018). The stratigraphic framework is important as it requires that of sub-ice-shelf sediment accumulated on the GZW foreset only after grounded ice backstepped to a new grounding location. There was not significant retreat of grounded and floating ice between the deposition of individual GZWs (McGlannan et al., 2017). It is not known whether the stacked GZWs represent phases of ice streaming with build-up alternated with phases of flow stoppage and/or some other process.

The stratigraphic framework also demonstrates that the ice shelf that covered the outer continental shelf eventually broke up during the deposition of GZW4 and this event is recorded by an event bed that overlies the sub-ice-shelf facies and underlies an open-marine diatom ooze (McGlannan et al., 2017). Three cores that penetrate the foreset of GZW7 do not exhibit the stratigraphic signature that is observed on the foresets of GZW1, -2 and -3. Instead, the homogenous till cored on the GZW7 foreset are directly overlain by open-marine sediments. These stratal relationships indicate that the ice shelf did not re-form over the outer continental shelf after its collapse (McGlannan et al., 2017). At the end of GZW7 deposition, the grounding line retreated more than 200 km toward Roosevelt Island. An ice shelf re-established over the middle continental shelf part of the Whales Deep Basin (i.e., to the south of the bathymetric saddle) as indicated by the presence of the fossiliferous sub-ice-shelf sediment on the middle continental shelf. The absence of seismically-resolvable GZWs to the south of the GZW7 indicates that the retreat of grounded ice towards Roosevelt Island was rapid (Bart et al., 2017a).

The distribution, orientation and resolution of the seismic lines are sufficient to map the large-volume of the stacked GZWs in the WDB. The subsurface control from seismic and seafloor geomorphologic evidence from multibeam bathymetric data indicates that the volumes of the first three GZWs (GZW1 to -3) are comparatively minor and that the bulk of the sediment belongs to the latter four GZWs (GZW4 to -7). The seismic velocity used to convert TWTT to depth (1700 m s^{-1}) is consistent with that derived for eastern Ross Sea by Cochrane et al. (1995). Our seismic correlation of a GZW unit in Joides Basin to the equivalent depth section in DSDP Leg 28 Site 273 is also consistent with a 1700 m s^{-1} velocity for time-depth conversion. A detailed map of the combined GZWs indicates that the total sediment volume is approximately 5.34×10^{11} m^3 (Bart et al., 2017a) (Table 1). A basic assumption of our calculations is that sediment transported by the BIS was deposited subglacially in the downstream reaches of the ice stream and/or at the grounding zone. In this context, the sediment yield reported in Table 1 column g represents the average erosion rate in mm/a averaged over the entire drainage area of the paleo BIS. The high erosion rates that occur after ice shelf breakup (e.g., 2.0 ±04 mm a^{-1}) probably include remobilization of recently deposited subglacial sediment. Our calculations do not include sediment that bypasses the grounding zone because those sediments cannot be seismically resolved. In the WDB, sub-ice-shelf sediments deposited coeval with the GZW are only ~1 m thick, much thinner than the average thickness of the GZW measured in tens of meters. Similar thin sub-ice-shelf deposits are reported in other Ross Sea basins (Domack et al., 1999; Anderson et al., 2013; Simkins et al., 2017; Prothero et al., 2108). If the sub-ice-shelf sediments have a uniform 1-m thickness in the area between the toe of the CGZW and the shelf edge, then the total volume would increase by only 1.4×10^{10} m^3 (i.e., less than 2% of seismically derived estimate of GZW sediment). Given this small volume, we do not include the inferred additional volume into our calculations.

We note that we would get very similar paleo ice stream velocities if we would use sediment flux parameters, $f = 0.5$, $h = 6$, from Alley et al. (1987). Hence, our results could not be used to prioritize the Christoffersen et al. (2010) sediment transport model over that of Alley et al. (1987). However, if we were to adopt the very low values of h from Engelhardt and Kamb (1998) and Hodson et al. (2016), we would obtain ice stream velocities in excess of 10,000 m a^{-1}, implying ice thinning rates of a few to several meters per year in the entire drainage basin of paleo-BIS over the 3200 years of GZW deposition. This is not feasible and we conclude, post factum, that the use of sediment flux parameters from either Christoffersen et al. (2010) or Alley et al. (1987) is a suitable choice for our analyses.

In fact, it is difficult to conceive that the sediment flux parameters, f and h, were very different for the paleo-BIS than they were for the case observed by Christoffersen et al. (2010)

and the one inferred by Alley et al. (1987). For instance, if paleo-BIS would be associated with twice as large $f \times h$ than we assumed in our baseline case, the paleo-ice stream drainage basin would not reach negative mass balance throughout the deposition of the GZW but would experience ice thickness growth (Figure 2). Conversely, if $f \times h$ were several times smaller than the baseline scenario (e.g., $f \times h$ = 1 m), ice velocities required to meet the observed sediment fluxes would cause basin-wide thinning rates of about 1.0 m a^{-1}, which could not be sustained either over the entire period of GZW deposition, or even the shorter, post-ice shelf breakup period without leading to premature abandonment of the Whales Deep GZW grounding line position on the outer continental shelf. Hence, it appears that the sediment flux parameters applicable to the paleo-BIS ice stream during its deglaciation did not differ by more than a factor of a few from the ones adopted in our calculations from Christoffersen et al. (2010) (or Alley et al., 1987). This suggests that Siple Coast ice streams are, and have been, efficient agents of sediment transport leading to erosion rates of a fraction of mm per year, at least during their active flow phases (Table 1).