Similarities in basal sliding between Greenland and Alpine Glaciers

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ABSTRACT

We document seasonal variations in basal sliding velocity along the terminal portion of a flowline in Western Greenland. Differential global positioning system observations from four stations show a strong annual cycle in ice velocity. Summer speedup (+36 to 56 %) and fall slowdown (-11 to 25 %) events at each station are interpreted as seasonal variations in basal sliding speed. We speculate that this Greenland flowline responds to seasonal melt input in a manner akin to alpine glaciers. We can efficiently characterize the annual velocity cycle using two Gaussian curves (representing summer and fall speeds) superimposed on the mean winter velocity. This simple model of seasonal velocity is dependent on five observable parameters that vary from site to site: maximum, minimum and winter velocities, as well as Julian Dates of maximum and minimum velocity. Based upon a projection of the spatial pattern of summer anomalous behavior, seasonal variations in basal sliding should not exist upstream of km 86.5.

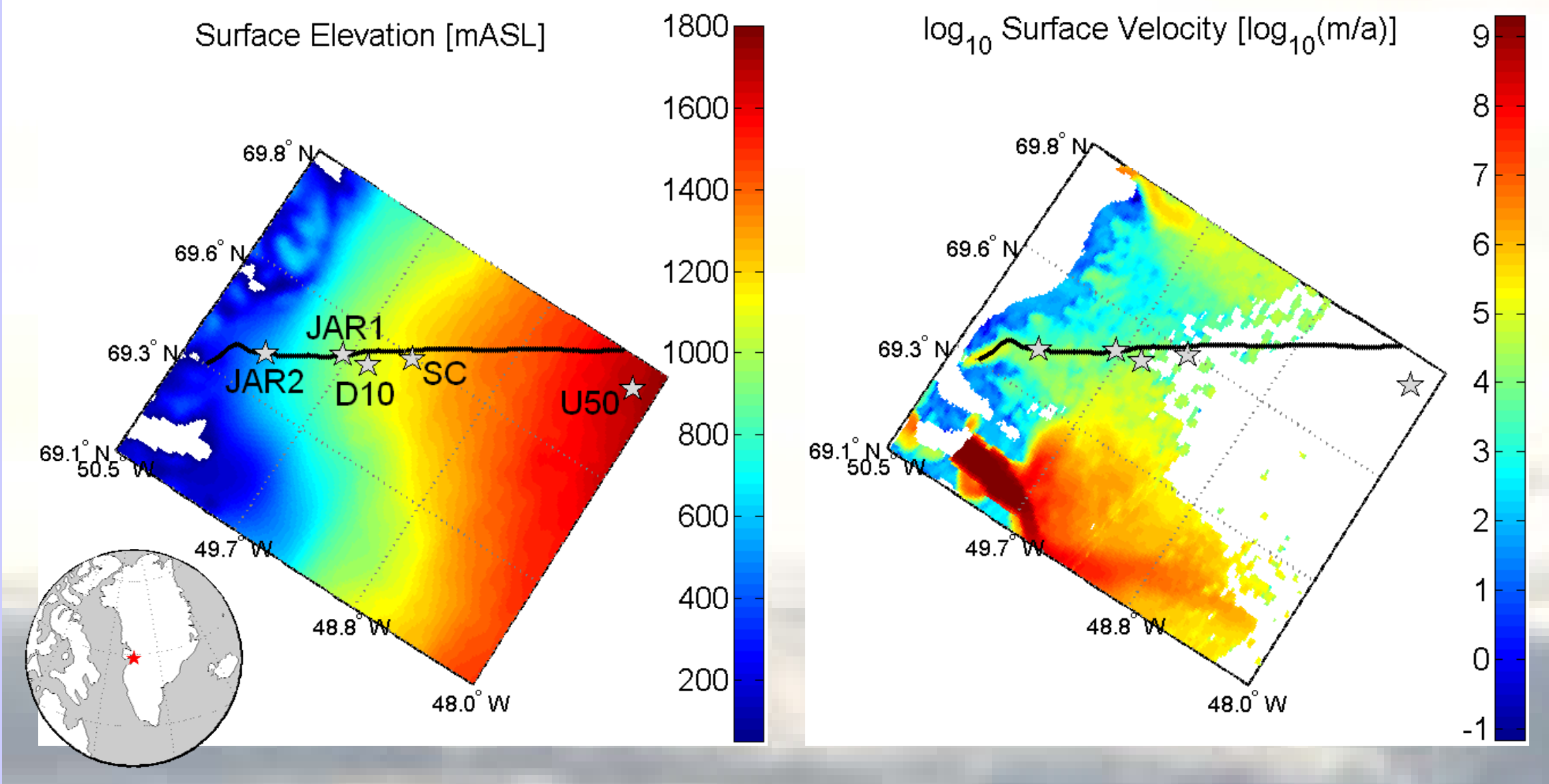

Figure 1 – *The terminal 100 km of the Sermeq Avannarleq ("Dead") Glacier flowline, just north of Jakobshavn Isbrae, overlaid on observed ice surface elevation (**Left**;* Scambos and Haran, *2002) and inferred ice surface velocity (**Right**;* Rignot and Kanagaratnam, *2006). The locations of the five DGPS stations are shown with stars (Table 1).*

STUDY SITE AND DATASET

Our study site is the terminal ~ 100 km of the Sermeq Avannarleq ("Dead") Glacier flowline in Western Greenland (Figure 1). This ~ 530 km flowline originates at the main ice divide of the Greenland Ice Sheet (71.54°N, 37.81°W). There are five differential global positioning system (DGPS) stations located along this flowline, known as JAR2, JAR1, Down 10, Swiss Camp and Up 50. Variations in basal sliding speed on synoptic time-scales have been previously observed at Swiss Camp (*Zwally et al*., 2002). Only DGPS data recorded after the year 2000 was used in this study. Station velocity values were calculated from differences in DGPS position over periods of between 20 and 60 days. The length of velocity datasets (DGPS sample size) varies between two years (Up 50) to nine years (Swiss Camp; Table 1). Although each station suffers data gaps over the nine-year study period, which makes direct inter-annual comparisons difficult, combining data from multiple years allows the intra-annual velocity cycle to be resolved at each station (Figure 2).

Table 2 – Characteristics of the five DGPS stations: latitude, longitude, elevation, position along flowline (terminus = 0 km), distance offset from flowline and DGPS sample size.

Station	Latitude (°N)	Longitude (°E)	Elevation (m)	Flowline Position (km)	Flowline Offset (km)	DGPS Sample Size (*n*)
JAR2	69.416	-50.083	531	14.5	0.23	1087
JAR1	69.496	-49.700	954	32.5	0.25	977
Down10	69.507	-49.544	1075	38.0	2.88	202
SC	69.563	-49.339	1152	47.5	1.93	1242
Up50	69.747	-48.136	1643	98.5	8.45	485

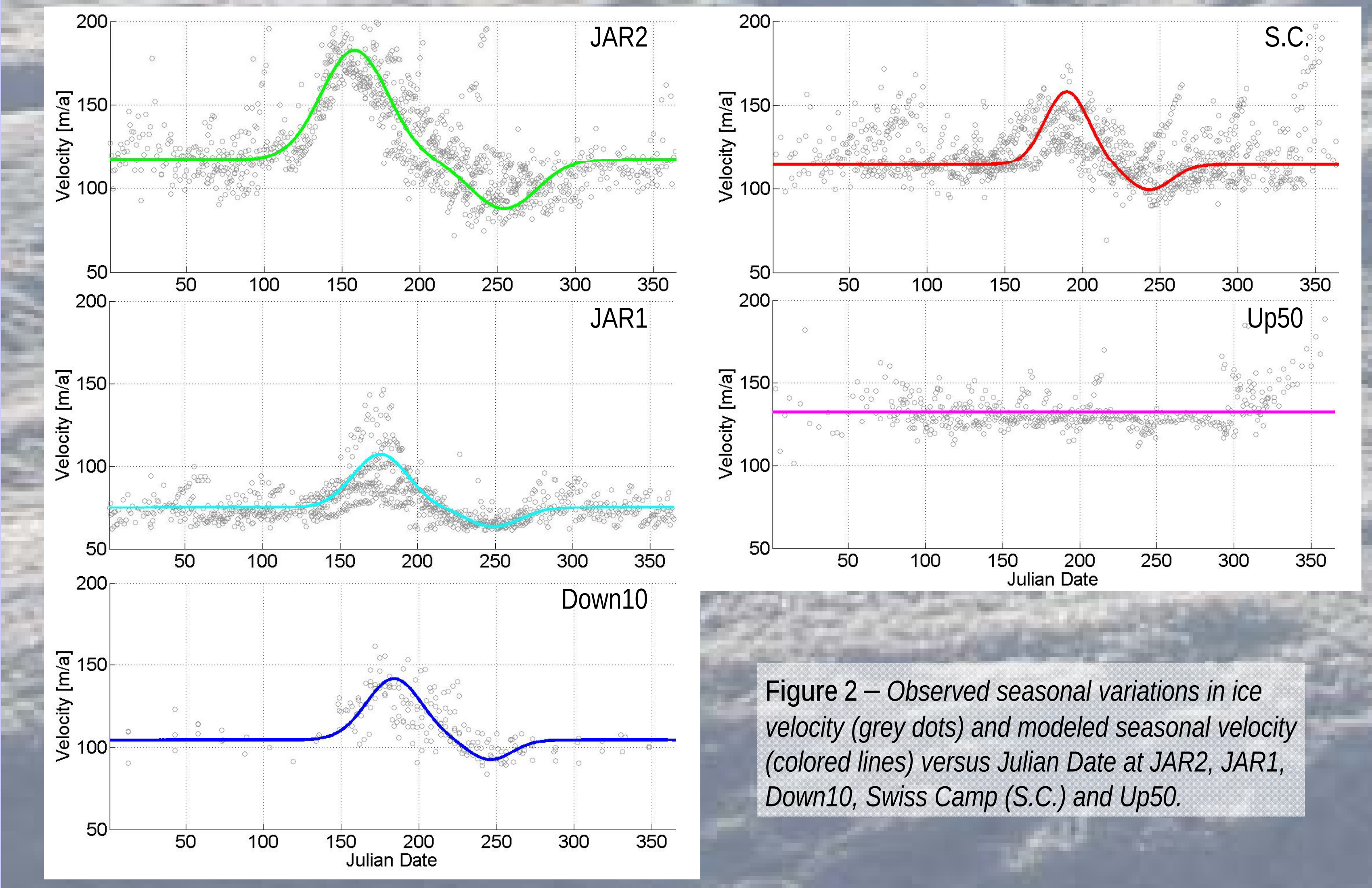

Figure 2 – *Observed seasonal variations in ice velocity (grey dots) and modeled seasonal velocity (colored lines) versus Julian Date at JAR2, JAR1, Down10, Swiss Camp (S.C.) and Up50.*

BASAL SLIDE PARAMETERIZATON

It is apparent that the seasonal velocity cycle at each station contains a summer speedup event and a fall slowdown event. We can efficiently characterize these annual velocity cycles using a simple model comprised of two Gaussian curves superimposed on the mean winter velocity, where one curve represents the summer speedup and the other curve represents the fall slowdown (Figure 2). Gaussian curves were chosen to represent the summer and fall velocity anomalies because the amplitude, width and timing of these curves can be independently parameterized. Thus, the annual surface velocity (U_s) at each station, on a given Julian Day (j), can be characterized by three terms:

$$U_s = U_w + (U_{max} - U_w)\cdot \exp\left[\left(\frac{j - J_{max}}{D_{max}}\right)^2\right] - (U_w - U_{min})\cdot \exp\left[\left(\frac{j - J_{min}}{D_{min}}\right)^2\right] \quad \text{Equation 1}$$

The winter velocity (U_w) was calculated as the mean of observed velocity values at a given station between JD 300 and 100. U_{max} represents the maximum summer velocity and was calculated as the mean of the highest quintile of velocity values at a given station between Julian Dates 100 and 250. Similarly, U_{min} represents the minimum fall velocity and was calculated as the mean of the lowest quintile of velocity values at a given station between JD 200 and 300. The remaining four parameters specify the timing and shape of the summer speedup and fall slowdown curves. J_{max} (J_{min}) represents the JD of summer maximum (fall minimum) velocity, while D_{max} (D_{min}) represents the duration of the summer (fall) velocity anomaly. These four parameters were evaluated for each station using a supervised optimization scheme (Table 2).

Table 2 – Basal slide parameters at each station: winter velocity, summer max velocity, fall min velocity, Julian Date of summer velocity max, duration of summer max, Julian Date of fall min and duration of fall min (Figure 2).

Station	Winter Velocity (m/a)	Summer Max. Vel. (m/a)	Fall Min. Vel. (m/a)	Summer Max. (JD)	Max. Duration (d)	Fall Min. (JD)	Min. Duration (d)
JAR2	117.4	182.6	87.7	158	30	254	30
JAR1	75.3	107.3	63.3	176	25	249	25
Down10	104.5	141.6	92.5	184	25	246	20
SC	114.3	158.0	99.2	190	20	243	20
Up50	132.5	-	-	-	-	-	-

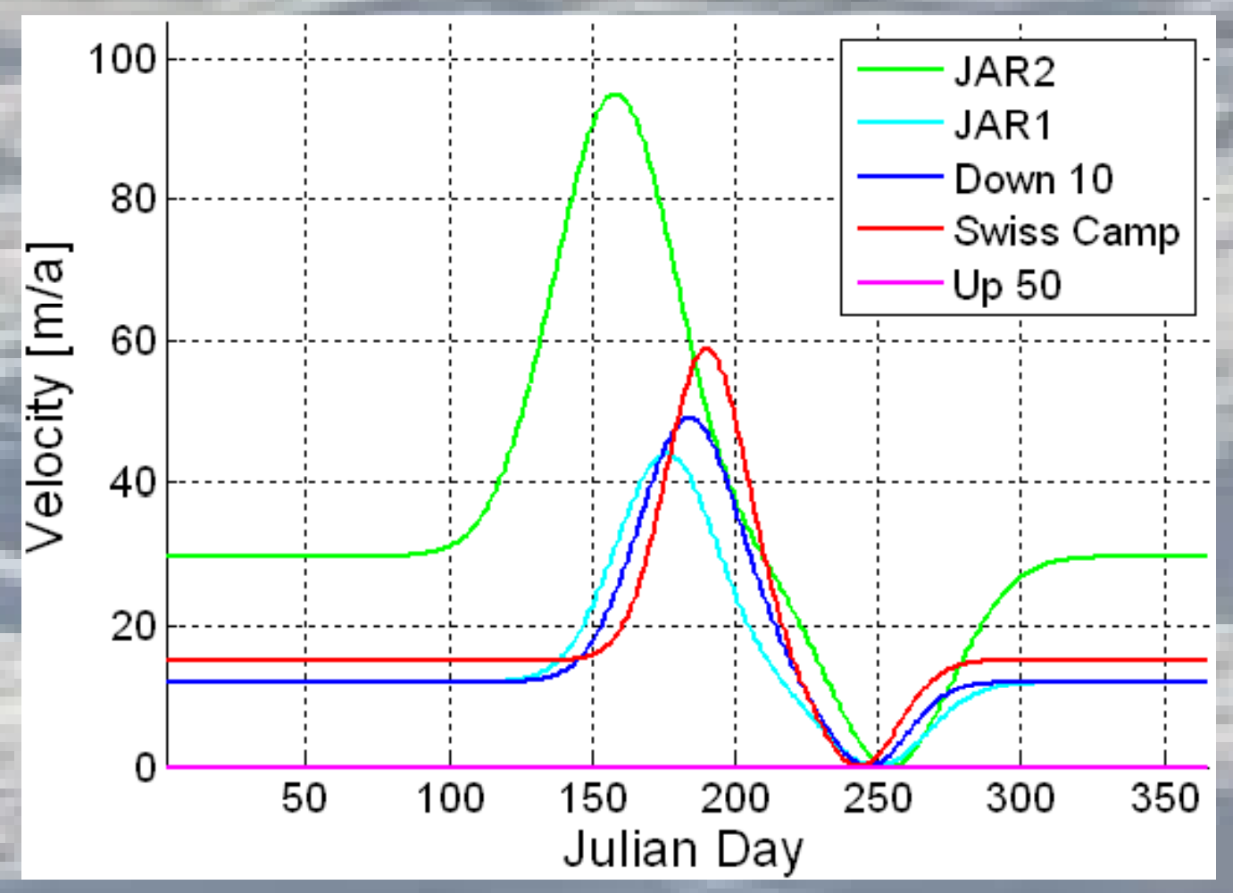

Figure 3 – *Modeled basal slide versus Julian Day at each of the five DGPS stations.*

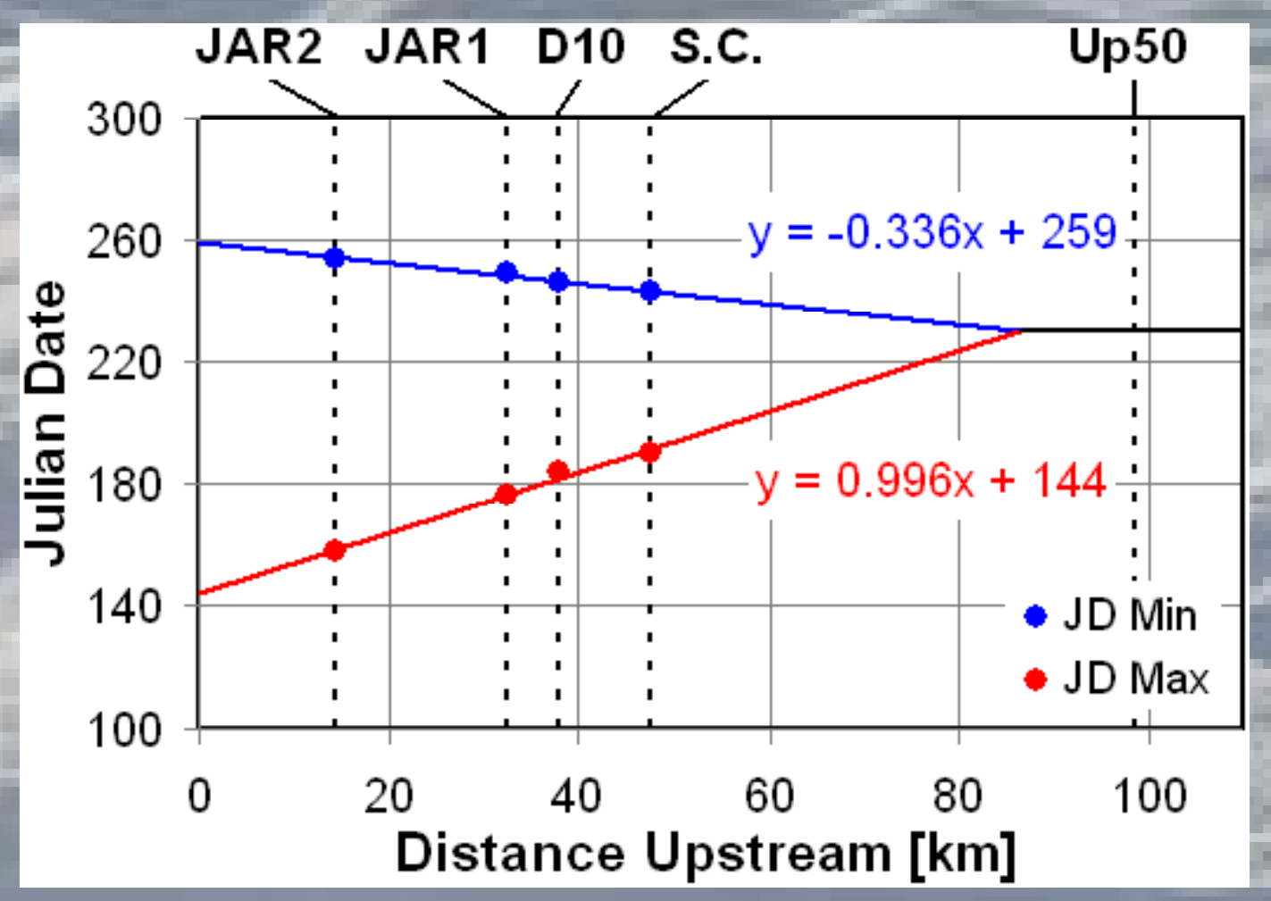

Figure 4 – *Julian Dates of max and min velocity versus distance upstream. The convergence suggests basal slide goes to zero at km 86.5.*

RESULTS

We interpret seasonal variations in surface ice velocity to represent seasonal variations in basal sliding speed. Thus, subtracting the fall minimum velocity from each station record yields the magnitude of seasonal basal slide (Figure 3). The DGPS observations at the four terminal stations reveal that the glacier moves at winter speeds until the beginning of a summer acceleration, which coincides with the onset of melt (Julian Day ~ 120), and peaks around JD ~ 177. The summer speedup event lasts ~ 50 days and exhibits velocities which are ~ 36 to 56 % above winter velocities. The summer event is immediately followed by deceleration to a fall minimum. This fall minimum at JD ~ 248 exhibits velocities which are ~ 11 to 25 % below winter velocities. Ice velocities increase again to winter speeds by JD 300.

A projection of the spatial pattern of dates of maximum and minimum speeds suggests that seasonal variations in basal sliding should not exist above ~ 86.5 km upstream from the terminus (Figure 4). This matches observations, as the four stations downglacier of km 86.5 exhibit annual velocity variations, while the one station upglacier of this point (Up50) does not (Figure 2). Knowing the seasonal basal slide history at each station allows the magnitude of basal slide to be interpolated through space and time along the terminal portion of the flowline (Figure 5).

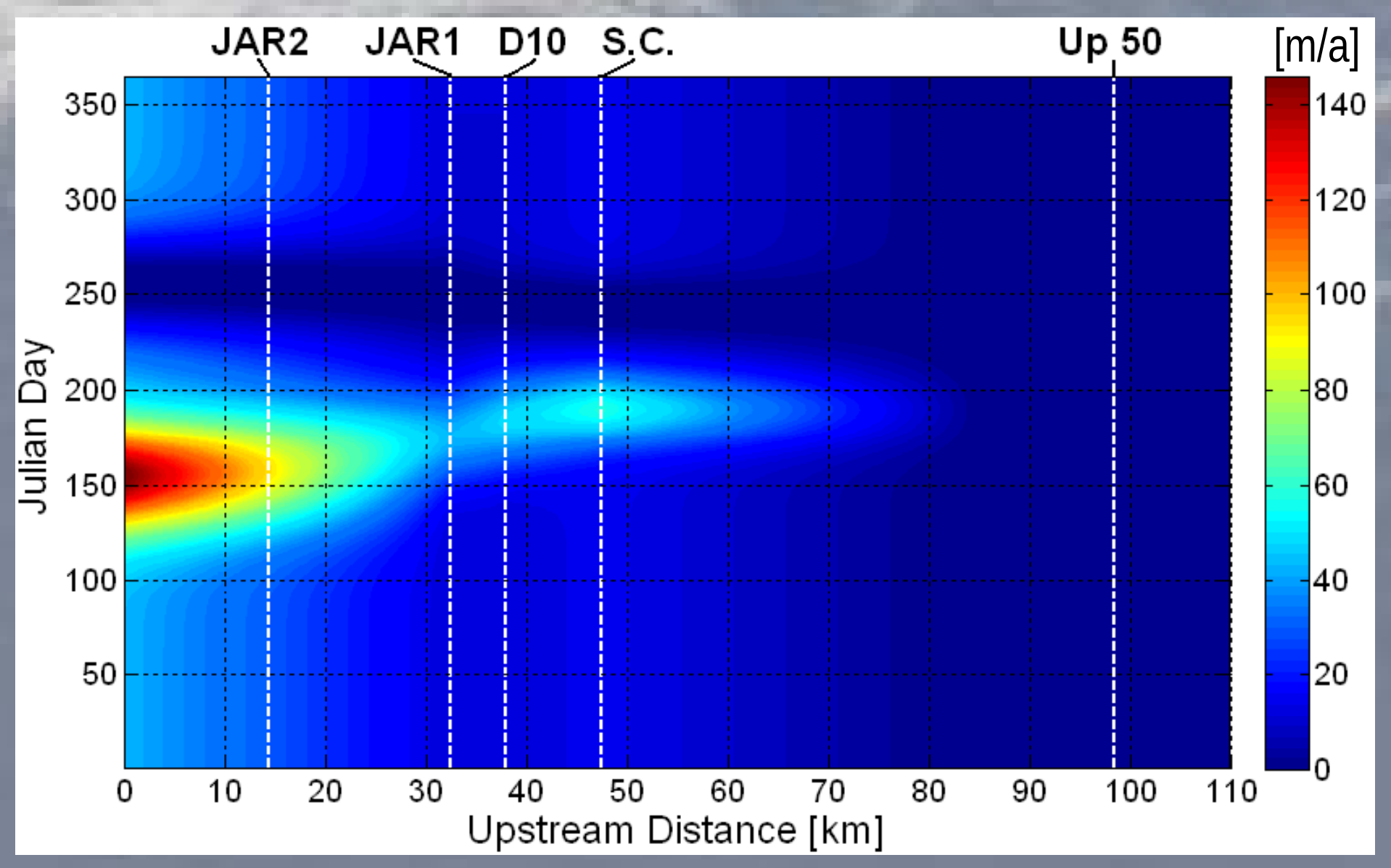

Figure 5 – *Modeled basal slide linearly interpolated through time (Julian Date) and space along the terminal 110 km of the flowline. The positions of the five DGPS stations are shown.*

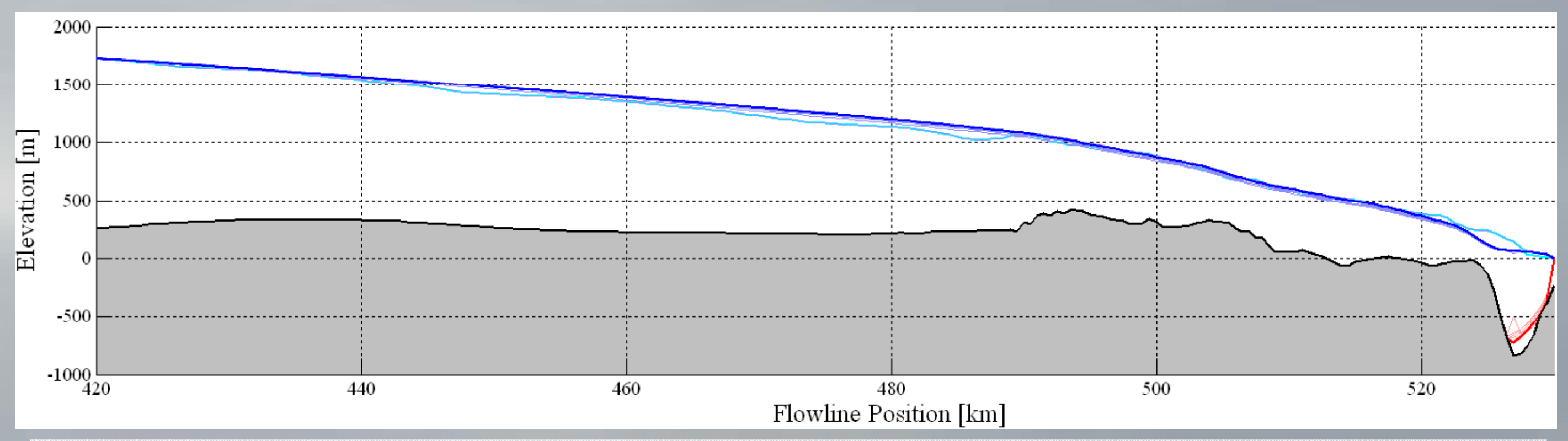

Figure 6 – *Observed ice surface elevation (cyan;* Scambos and Haran*, 2002) and modeled ice surface (blue) and floating bottom (red) elevation along the terminal portion of the flowline after a 505-year simulation. Observed bedrock topography is shown is grey (*Bamber et al.*, 2001;* Plummer et al.*, 2008).*

FLOWLINE MODEL

The observed spatial and temporal distribution of basal slide was used as a basal boundary condition in a flowline model. Important components of the flowline model are: (i) longitudinal coupling (*van der Veen*, 1987), (ii) a temperature-dependent flow law parameter (*Huybrechts et al*., 1991), (iii) a Wisconsinan flow enhancement factor of three (*Reeh*, 1985; *Paterson*, 1991), (iv) a floating tidewater terminus with equilibrium calving flux and (v) a semi-implicit numerical method that allows time-steps of up to 25 years (Figure 6). The model demonstrates that the peak magnitude of seasonal basal slide is approximately equal to that of internal deformation in the lower reaches of the flowline (Figure 7). Although modeled surface ice velocities compare reasonably well with observed DGPS velocities, modeled surface ice velocities (as well as DGPS velocities) are consistently higher than those inferred by InSAR interferometry (*Rignot and Kanagaratnam*, 2006).

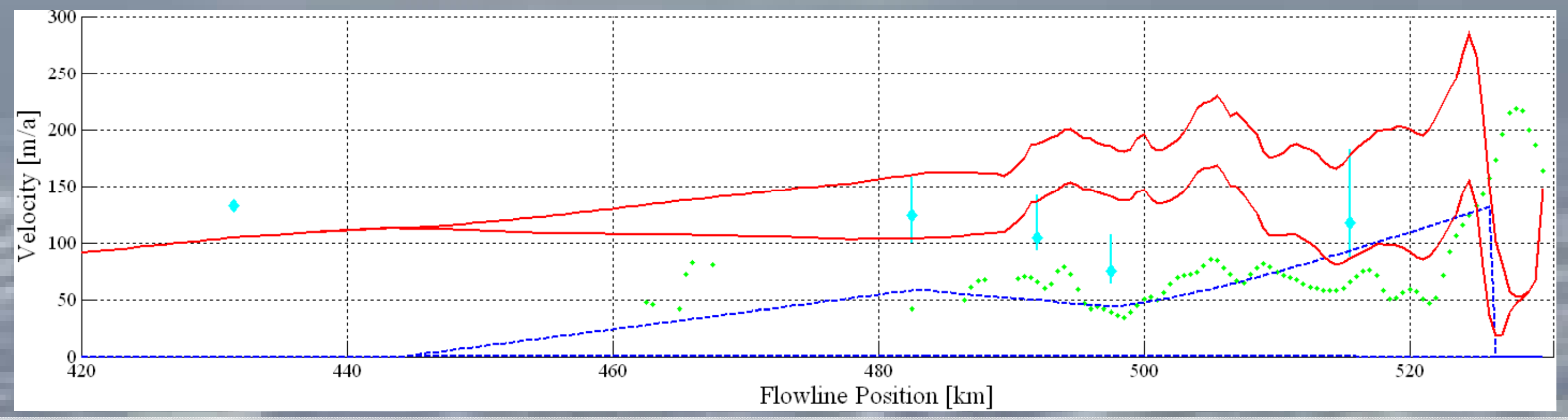

Figure 7 – *Seasonal max and min modeled ice surface velocity (red) and seasonal max and min prescribed basal slide (blue dashed; Figure 5), along the terminal portion of the flowline after a 505-year simulation. Observed summer, fall and winter velocities are shown at the five DGPS stations (cyan). InSAR-derived interferometric velocity along the flowline is also shown (green dots; Figure 1;* Rignot and Kanagaratnam, *2006).*

Figure 8 – *Servicing the JAR2 (**Left**) and Down 10 (**Right**) autonomous DGPS stations in the spring of 2008.*

INTERPRETATION AND CONCLUSIONS

We speculate that the terminal portion of this flowline of the Greenland Ice Sheet responds to seasonal melt input in a manner akin to the spring speedup events on alpine glaciers, whereby meltwater-induced pressurization of the subglacial system early in the melt season is followed by deceleration due to the establishment of efficient drainage and decreasing meltwater input towards the end of the melt season (*Anderson et al*., 2004; *Bartholomaus et al*., 2008). The sensitivity of both the magnitude of seasonal basal sliding, and the duration of the sliding season to distance along flowline, suggests that a small inland migration in the location of basal sliding onset should result in a non-linear increase in annual total ice discharge to the ice sheet margin.

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Background Photo: Jakobshavn Ablation Region, August 2009