The 2007 Bering Strait Oceanic Heat Flux and anomalous Arctic Sea-ice Retreat

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Abstract*:* To illuminate the role of Pacific Waters in the 2007 Arctic sea-ice retreat, we use observational data to estimate Bering Strait volume and heat transports from 1991 to 2007. In 2007, both annual mean transport and temperatures are at record-length highs. Heat fluxes increase from 2001 to a 2007 maximum, 5-6x10^{20}J/yr. This is twice the 2001 heat flux, comparable to the annual shortwave radiative flux into the Chukchi Sea, and enough to melt 1/3rd of the 2007 seasonal Arctic sea-ice loss. We suggest the Bering Strait inflow influences sea-ice by providing a trigger for the onset of solar-driven melt, a conduit for oceanic heat into the Arctic, and (due to long transit times) a subsurface heat source within the Arctic in winter. The substantial interannual variability reflects temperature and transport changes, the latter (especially recently) being significantly affected by variability (> 0.2Sv equivalent) in the Pacific-Arctic pressure-head driving the flow.

1. Introduction

How significant was heat flux from Pacific Waters (PW) in the extreme 2007 Arctic sea-ice retreat? Woodgate *et al.* [2006] estimated that Bering Strait oceanic heat fluxes from the early 2000s were substantial, 2-4x10^{20}J/yr, and that the extra heat input from 2001 to 2004 was enough to melt 640,000km^2 of 1m thick ice, comparable to the change in summer ice-extent (700,000km^2) between 2001 and 2004 [*Stroeve et al.*, 2005]. Shimada *et al.* [2006] propose a positive feedback where heat carried by northward flowing PW weakens the ice-pack thereby promoting more sea-ice motion in response to wind, which in turn enhances the wind-driving of PW into the Arctic. Observationally-based estimates of Bering Strait fluxes are essential to quantitatively test such hypotheses and to evaluate studies of Arctic change.

2. Quantification of Bering Strait fluxes

2.1 Data Sources

The Bering Strait - ~ 85km wide, 50m deep and divided into two channels by the Diomede Islands - is the only oceanic gateway between the Pacific and the Arctic Oceans. Year-round near-bottom oceanographic moorings (Figure 1a) have been deployed in the region almost continuously since 1990, generally at three locations – one at a mid-strait site (A3, ~ 66.3ºN 169ºW) ~ 60km north of the Diomede Islands, and one at the centre of each channel (A2, eastern channel; A1, western channel).

Although the annual mean transport through the strait is northward (~ 0.8Sv), there is substantial higher frequency variability: 0 to 1.5Sv northward monthly; 2Sv southward to 3Sv northward daily [Figure 2a and *Woodgate et al.*, 2005b]. It appears that the velocity at site A3 correlates well with the total volume transport through the strait - mooring and ship-based data show high coherence in velocity at all sites in the strait region [*Woodgate et al.*, 2005b], with velocities being higher in the Alaskan Coastal Current (ACC), a warm, fresh, ~ 0.1Sv current present along the Alaskan coast from midsummer until ~ December [Figure 1a, and *Woodgate and Aagaard*, 2005]. This strait-scale coherence is expected since the suspected forcing for the flow – a pressure-head difference between the Pacific and the Arctic, mediated by local wind-driven effects – is on length-scales greater than the strait width [see *Woodgate et al.*, 2005b, for discussion]. Lacking year-round velocity cross-sections, we assume the flow field is homogeneous and barotropic and use a cross-section area of $4.25 \times 10^6 m^2$ to convert A3 velocity to transport. This probably underestimates the transport (as it neglects the ACC), and may introduce systematic errors (likely < 20%).

Temperatures in the strait vary substantially, especially seasonally – from freezing in winter to 4-10ºC (near-bottom/surface data) in late summer [Figure 2c and *Woodgate et al.*, 2005a]. Other than in winter, there is a cross-strait temperature gradient with warmer temperatures in the east. Yet mooring data suggest that site A3, being central to the strait, yields a reasonable average of the mean near-bottom temperatures (good to ~ 0.1ºC/0.5ºC in annual/monthly means [*Woodgate et al.*, 2007]). Thus, we use site A3 to estimate mean near-bottom temperatures. Due to ice-keels, most in situ measurements in the strait are deep, 10m above the bottom, and thus underestimate water column mean temperatures since summer/autumn CTD sections indicate a 10-20m thick surface layer, 1-2ºC warmer than below. To estimate temperatures in this layer, we use 7-day averages of AVHRR [*Vazquez et al.*, 1998] Sea Surface Temperature (SST) from within ~ 40km of A3 (spatial and temporal averaging selected to reduce time-series gaps in this typically cloudy region). We estimate this assumption is good to ~ 1ºC, considering stated errors and a comparison with available coincident in situ CTD data. Since we seek the heat available to melt sea-ice in the Arctic, our temperature reference is -1.9ºC, the freezing point of Bering Strait waters, reflecting that these waters lose most of their heat before leaving the Arctic [*Steele et al.*, 2004].

Using mooring data alone (and assuming a barotropic, homogeneous water column), we compute heat fluxes at the finest time resolution available (typically hourly). Then we estimate a simplistic stratification correction, using SST and two nominal upper layer thicknesses of 10 and 20m (Figure 2e). Overall, uncertainties are of order 0.1Sv, 0.8×10^{20}J/yr. To include the ACC, we add ~ 0.1Sv, and ~1×10^{20}J/yr to the northward fluxes [*Woodgate et al.*, 2006]. Since A3 data are incomplete before 1998, we also present results from eastern channel (A2) temperature and velocity data, to allow comparison back to 1991.

2.2 Results from 1991 to 2007

Annual mean values (Figure 2) are calculated for calendar years. For volume transport, the interannual variability (0.6-1.0Sv) masks any long term trend (Figure 2b). The record low (0.6Sv, 2001) reflects weaker northward flow in many periods of the year, while the second lowest (0.7Sv, 2005) is mostly due to anomalous southward flow in late 2005 (Figure 3). The highest transports (~ 1Sv, 2004 and 2007) have strong northward flows in winter and for most of the summer.

The most dramatic change in A3 (i.e., near-bottom) annual mean temperatures is a ~1ºC increase between 2001 and 2002 (Figure 2d). Although 1993 was possibly warm (A2 data), A3 means from 2002 -2007 are 0.1-0.4ºC, compared to -0.2 to -0.5ºC pre 2002. There is no obvious match between high transports and high temperatures, although it proves important that both are high in 2007. The SSTs, also with a record high in 2007 (Figure 2d), show slightly larger interannual variability than near-bottom temperatures.

Annual mean heat fluxes (Figure 2e) increase almost monotonically from 2001 to 2007. Heat flux variations are driven almost equally by changes in volume transport and in temperature, and the highest heat flux years (2004 and 2007) are those where both transport and temperature are high. Year 2007 yields a clear record-length maximum, estimated at $3.5x10^{20}$J/yr from A3 data alone and at 4-$4.7x10^{20}$J/yr including a 10-20m surface layer. Adding ~ $1x10^{20}$J/yr for the ACC yields a total heat flux of 5-$5.7x10^{20}$J/yr. This is almost a doubling of the total 2001 heat flux (~ 2.6-$2.9x10^{20}$J/yr), and $1x10^{20}$J/yr greater than the previous high in 2004 (~ 4.3-$4.8x10^{20}$J/yr).

3. Implications for Arctic Sea-ice retreat

How relevant is this amount of heat (3-$6x10^{20}$J/yr, i.e., 10-20TW) in the Arctic?

This much heat could melt 1-$2x10^6km^2$/yr of 1m thick ice, maybe 1/3rd of annual arctic sea-ice retreat and comparable to interannual variability in the September ice extent – winter extent is ~ $10x10^6km^2$; the 2006 and 2007 September minima were 6 and $4x10^6km^2$ respectively (National Snow and Ice Data Center data).

Pacific Waters (PW) are found over roughly half the Arctic Ocean. Averaged over that area (~$5x10^6km^2$), the Bering Strait heat flux is 2-$4W/m^2$, a significant fraction of Arctic annual mean net surface heat fluxes (-2 to $10W/m^2$, ERA-40 atmospheric reanalysis [Figure 5 of *Serreze et al.*, 2007]).

The Bering Strait heat flux is also comparable to the solar input to the Chukchi Sea, ~ $4x10^{20}$J/yr (~ 1300 $MJm^{-2}yr^{-1}$, 1998-2007 range, *Perovich et al.,* [2007 updated], Chukchi Sea area ~$350x10^3km^2$).

In fact, the Bering Strait heat flux is surprisingly large for its net volume. Although the Fram Strait inflow is about 10 times greater in volume, its estimated net heat input to the Arctic is ~ 30-50TW [*Schauer et al.,* 2008], only about 3 times our Bering Strait estimate.

Thus, purely in terms of heat, the Bering Strait contribution is large enough to be a significant player in sea-ice retreat. But other factors are also important, viz., the timing of the delivery of this heat to the Arctic; the volume throughflow itself, which may carry heat and ice northward; and (since we seek explanations for interannual change) the magnitude of interannual variability of these properties.

3.1 Bering Strait throughflow as a trigger to start the seasonal melt back of ice

The ice-albedo feedback – less ice allows more shortwave energy to be absorbed by the water, warming the water and thus melting more ice – is often cited as the main cause of sea-ice retreat. However, other than in leads and surface melt-ponds, the albedo feedback generally only takes over once a region of open water has been created.

In 2007, a small open water area appeared south of the strait in early April, when the water flow was southward. When the flow turned north about 2 weeks later, open water appeared north of the strait, coincident with slight (0.1ºC above freezing) near-bottom warming at A3. Through May the flow remained northwards and the sea-ice retreat occurred in patterns consistent with known Chukchi flow pathways via Herald Valley, the Central Channel and Barrow Canyon [*Weingartner et al.*, 1998; *Woodgate et al.,* 2005b] and, by the end of June, these pathways appear dramatically in the ice-melt pattern (Figure 1b).

Even though initial opening may be due to ice-motion and/or melt, the ice retreat pattern strongly suggests that water pathways through the Chukchi are acting as a conduit of heat (gained either locally or further south) into the Arctic and, furthermore, are providing the initial heat (and/or motion) necessary to open up ice in the Arctic Basin. Conservation of volume dictates an increase of Bering Strait transport must be rapidly compensated with increased outflow to the Arctic Basin. (A 1Sv imbalance would change the Chukchi sea-level by 1m in about 4 days). Thus, strong northward transport in the Bering Strait (usual in summer, Figure 3) must correspond to strong northward flow of waters into the Arctic Basin.

3.2 Pacific Waters as a time-delayed source of subsurface heat in the Arctic

In terms of delivering heat to the north, we must also consider the time taken to transit the Chukchi, estimated at several months by *Woodgate et al.*, [2005b]. Mooring data from 2002-2004 at ~73.5ºN 166ºW (http://www.eol.ucar.edu/projects/sbi/mooring.shtml) indicate that near-bottom seasonally warm waters arrive/leave later in the northern Chukchi (December/February) than in the Bering Strait (May/December). Thus, subsurface heat – albeit modest (maximum 2ºC, typically -0.5 to 0ºC) – is present under the surface ice cover and is provided to the Arctic in the middle of winter.

Within the western Arctic Ocean, the PW form a subsurface temperature maximum that weakens away from the Chukchi. The heat stored in PW is ~ 40-140MJ/m^2 depending on location [*Steele et al.*, 2004]. Since the PW have cooled almost to freezing when they leave the Arctic, this heat must be lost somewhere, likely either to melt ice (100MJ/m^2 could melt 0.3m of ice) or to the atmosphere via leads. Thus again, the PW heat appears to be a modest, but significant quantity.

Note this implies a memory to the system. The heat supplied to the Arctic Basin in January 2008 relates to the summer 2007 Bering Strait heat flux. Similarly, the subsurface PW temperature maximum in the western Arctic is almost ubiquitously the remnant of an earlier summer's heating in the shelf seas, and thus significant interannual variability in heat input may have a delayed effect on Arctic sea-ice.

3.3 Interannual variability of the Pacific influence

One of our goals is to inform the discussion of interannual sea-ice variability in the Arctic. How large is the interannual variability in Bering Strait fluxes?

The 2007 transport is more than 1.5 times the 2001 transport (1.0 Sv compared to 0.6 Sv). Since the Bering Strait inflow is the largest input to the Chukchi, this implies significant change in resident time of waters in the Chukchi (~5.5 months, compared to 9 months, using Chukchi Sea volume, 14x10^{3}km^3, divided by transport as a simplistic estimate). This has many possible implications for the physics and biogeochemistry of the Chukchi, especially since Bering Strait annual mean temperatures also rise from -0.4ºC in 2001 to +0.4ºC in 2007.

The Bering Strait heat flux doubles between 2001 and 2007 (2-3x10^{20}J/yr to 5-6x10^{20}J/yr). Is this variability large compared with other sources? Using daily data (courtesy of B. Light) from the Chukchi south of 71ºN as per *Perovich et al., [2007]*, we estimate the annual shortwave input to the Chukchi between 1998 and 2007 varies between ~3 and 4.5x10^{20}J/yr. Thus, although both are relevant in terms of total heat input to the Arctic Basin, the Bering Strait heat flux variability appears to be slightly larger.

The Bering Strait heat flux variability is likely so high as it reflects changes both in volume transport and in temperature. What sets the variability in these two terms? There is little understanding of Bering Strait temperatures, which are presumably a complex result of atmospheric and solar effects on the northward moving waters of the Bering Sea. For transport, however, it is generally thought that the flow has two main drivers – a Pacific-Arctic pressure-head (of debated origin, often assumed constant) and local wind forcing [see *Woodgate et al.*, 2005b for discussion].

Assuming most simply "transport = PH + C x Wvel" where PH represents transport due to the pressure-head term, C is an unknown constant, and Wvel is the wind component with highest correlation to the flow (~ 330º), we seek PH and the constant, C, by a linear best fit of 6-hourly NCEP (National Centers for Environmental Prediction) 10m wind to estimated A3 transports in 1-year segments of data. Woodgate *et al.* [2006] tentatively related interannual transport variability to changes in the annual mean northward winds, admitting that the time-series was inconclusively short. The extended analysis (Figure 2f, wind velocity converted to equivalent transport assuming a 2001-2007 mean for "C') suggests, overall, the wind acts to slow the flow by between 0.4 and 0.2 Sv, although exact numbers vary strongly with wind point used. However, interestingly, the interannual variability in the PH transport is found to be comparable in magnitude to the wind-driven variability, with the total PH transport greater in 2007 (Figure 2g). This suggests that recent changes in the Bering Strait transport relate not just to changes in the local wind, but also to significant variability (> 0.2Sv equivalent) in the Pacific-Arctic pressure difference driving the flow [as suggested by *Woodgate et al.*, 2005b].

5. Conclusions

Using year-round data from in situ moorings and satellite-sensed sea surface temperatures, we quantify oceanic fluxes of volume and heat from the Pacific to the Arctic via the Bering Strait between 1991 and 2007 with special focus on 1998 to 2007. We find heat flux increases almost monotonically from 2001 to 2007. Reflecting both high volume transports and high temperatures, the estimated 2007 heat flux was the greatest recorded to date, 5-6x10^{20}J/yr (range reflecting uncertainty in depth of the summer surface layer). This is almost a doubling of the total 2001 heat flux and somewhat greater than the incoming shortwave solar input into the Chukchi Sea. Moreover, the interannual variability in the Bering Strait heat flux is slightly larger than that of shortwave solar input to the Chukchi.

We suggest that PW flow (both in terms of the transport of ice and of far-field and locally gained heat) acts initially as a trigger for the onset of the seasonal melt back of ice, and subsequently may drive a year-round modest thinning of the western Arctic ice, as it feeds a subsurface temperature maximum under the ice-pack in winter. Factors such as large interannual variability in heat flux, timing of the delivery of heat to the Arctic Ocean, and volume flux are also important. Thus, overall, the Bering Strait's leverage on the Arctic system may be greater than its comparatively small volume may suggest.

The data suggest that change in volume flux, which drives about half of the change in heat flux, is due not just to varying winds but also to significant (> 0.2 Sv equivalent) variation in the large-scale pressure-head forcing between the Pacific and the Arctic. This suggests that estimating the flow from wind-data alone will underestimate flux variability. Similarly, sea-surface temperatures show more interannual variability than the temperatures of the bulk of the water column, suggesting that flux estimates that are tightly coupled to SST may overestimate flux variability.

The oceanic heat that passes through the Bering Strait is solar in origin. Although in terms of delivery of heat to the high Arctic we must also consider solar input in the Chukchi, measuring the oceanic heat entering through the Bering Strait gives a useful boundary condition and is undoubtedly simpler and more accurate than assessing the heat budget over the Bering Sea to the south.

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FIGURES

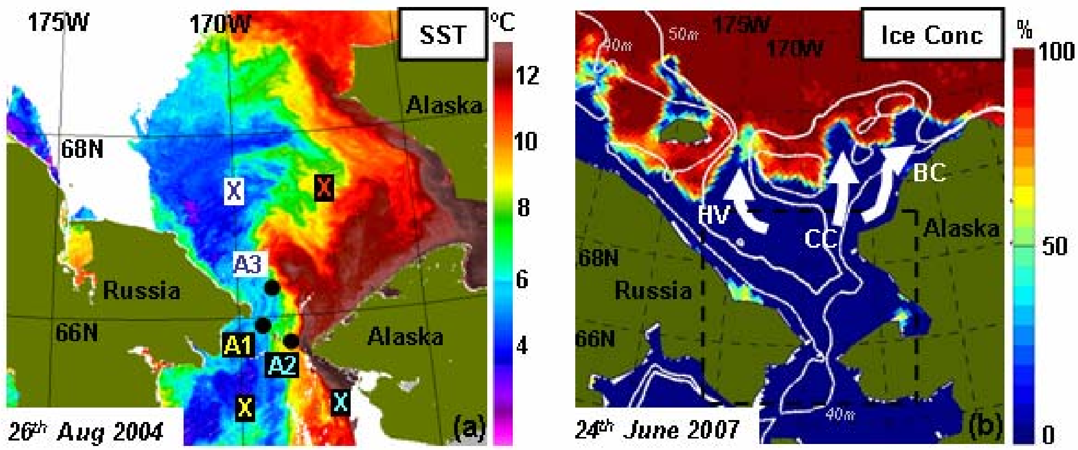

Figure 1: (a) Bering Strait summer MODIS Sea Surface Temperature (SST), marking moorings (black dots) and NCEP wind points (x, colors as per Figure 2f). **(b)** Chukchi Sea AMSR-E sea-ice concentration, with schematic topography. White arrows mark the three main water pathways via HV, Herald Valley; CC, Central Channel; and BC, Barrow Canyon. Black box marks region of (a).

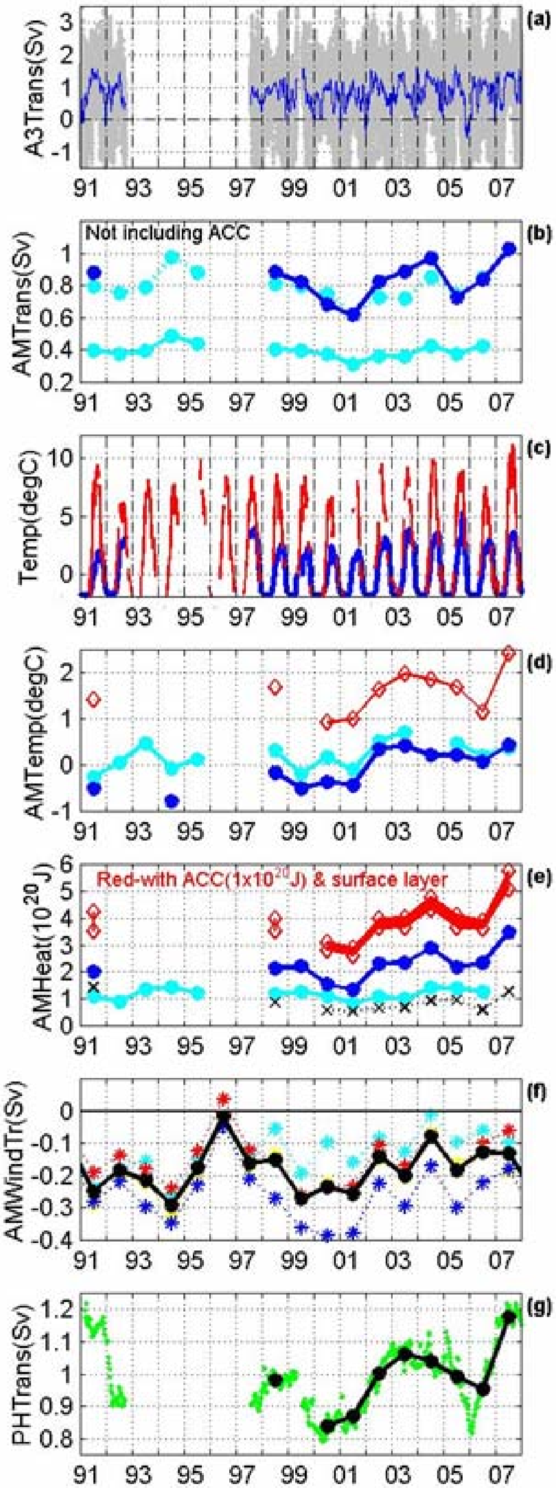

Figure 2: Bering Strait time-series from 1991-2007. Uncertainties are ~ 0.1Sv, 0.8x10^{20}J/yr. See text for possible systematic errors.

(a) Volume transport, from A3 data alone – grey, hourly; blue, 30-day smoothed.

(b) Annual mean (AM) transports (not including the Alaskan Coastal Current (ACC), ~ 0.1 Sv) – blue, total from A3; dashed cyan, total from A2; solid cyan, eastern channel transport only from A2.

(c) Temperatures – blue, 30-day smoothed A3 near-bottom; red, 7-day average AVHRR-SST.

(d) AM temperature– blue, A3 near-bottom; cyan, A2 near-bottom; red, SST near A3.

(e) AM heat fluxes – cyan, from A2 for eastern channel only; blue, total from A3 data only; red area, total including ACC correction (~ 1x10^{20}J/yr) and a 10m (lower bound) or 20m (upper bound) surface layer. Black crosses, amount of heat added by a 20m surface layer.

(f) Transport attributable to AM NCEP wind (heading 330°, i.e., ~ northwestward) - black, average over 4 nearest points (marked on Figure 1a) i.e., blue and red, ~150km north of the strait (67.5°N,170°W and 167.5°W); yellow and cyan, ~ 100km south of the strait (65°N, 170°W and 167.5°W).

(g) Transport attributable to the pressure-head term from annual fits – green, weekly data; black, annual mean.

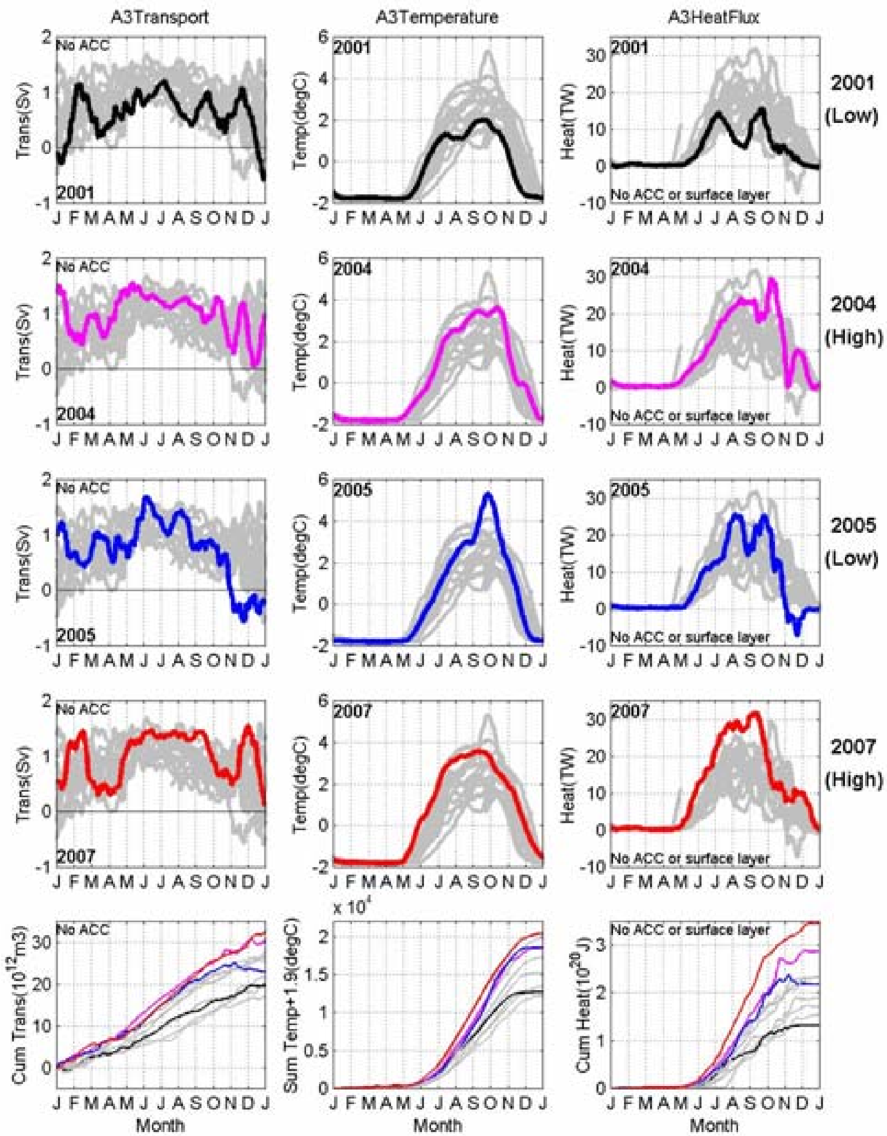

Figure 3: For extreme heat flux years (rows – lows: 2001, 2005; highs: 2004, 2007), 30-day smoothed A3-transports (left column), temperatures (middle column) and heat flux (right column), compared to the entire data set from 1991-2007 (grey). Last row: Cumulative sums of the above, colors indicating year as per upper panels.