Arctic Report Card 2016

Persistent warming trend and loss of sea ice are triggering extensive Arctic changes

2016 Headlines

Persistent warming trend and loss of sea ice are triggering extensive Arctic changes.

Observations in 2016 showed a continuation of long-term Arctic warming trends which reveals the interdependency of physical and biological Arctic systems, contributing to a growing recognition that the Arctic is an integral part of the globe, and increasing the need for comprehensive communication of Arctic change to diverse user audiences.

Video

Highlights

- The average **surface air temperature** for the year ending September 2016 is by far the highest since 1900, and new monthly record highs were recorded for January, February, October and November 2016.
- After only modest changes from 2013-2015, **minimum sea ice extent** at the end of summer 2016 tied with 2007 for the second lowest in the satellite record, which started in 1979.
- **Spring snow cover extent** in the North American Arctic was the lowest in the satellite record, which started in 1967.
- In 37 years of **Greenland ice sheet** observations, only one year had earlier onset of spring melting than 2016.
- The Arctic Ocean is especially prone to **ocean acidification**, due to water temperatures that are colder than those further south. The short Arctic food chain leaves Arctic marine ecosystems vulnerable to ocean acidification events.
- Thawing permafrost releases **carbon** into the atmosphere, whereas greening tundra absorbs atmospheric carbon. Overall, tundra is presently releasing net carbon into the atmosphere.
- Small Arctic mammals, such as **shrews**, and their parasites, serve as indicators for present and historical environmental variability. Newly acquired parasites indicate northward shifts of sub-Arctic species and increases in Arctic biodiversity.

CAFF Conservation of Arctic Flora and Fauna

December 2016

www.arctic.noaa.gov/Report-Card

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Executive Summary

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The Arctic Report Card (www.arctic.noaa.gov/Report-Card/) considers a range of environmental observations throughout the Arctic, and is updated annually. As in previous years, the 2016 update to the Arctic Report Card highlights the changes that continue to occur in, and among, the physical and biological components of the Arctic environmental system.

Arctic air temperatures continue to increase at double the rate of the global temperature increase. The average annual surface air temperature anomaly (+2.0° C relative to the 1981-2010 baseline) over land north of 60° N between October 2015 and September 2016 was by far the highest in the observational record beginning in 1900. This represents a 3.5° C increase since the beginning of the 20th Century. Autumn, spring and winter showed extensive positive average air temperature anomalies across the central Arctic, primarily due to southerly winds moving warm air into the Arctic from mid-latitudes. Winter air temperatures greatly exceeded the previous record, with several locations showing January temperature more than 8° C above the norm. Contrary to conditions in much of the previous decade, neutral to cold temperature anomalies occurred across the central Arctic Ocean in summer 2016.

The relatively cool summer air temperatures over the Arctic Ocean created a condition which did not support rapid summer sea ice loss. After experiencing the lowest winter maximum ice extent in the satellite record (1979-2016)—which occurred in March and was 7% below the 1981-2010 average—many anticipated a record summer minimum extent. Though no new record was set, the September 2016 Arctic sea ice minimum extent tied with 2007 for the second lowest value in the satellite record, at 33% lower than the 1981-2010 average. The sea ice cover continues to be relatively young and thin. In March 2016, multiyear ice (more than 1 year old) and first-year ice were 22% and 78% of the ice cover, respectively, compared to 45% and 55% in 1985.

As the sea ice retreats more extensively in the summer, previously ice-covered water is exposed to more solar radiation. As a result, sea surface temperature (SST) and upper ocean temperatures are increasing throughout much of the Arctic Ocean and adjacent seas. Also contributing to the increase in SST are regional air temperatures and influx of warmer water from the North Atlantic and Pacific Oceans. The Chukchi Sea, northwest of Alaska, and eastern Baffin Bay, off west Greenland, have the largest warming trends: ~0.5° C per decade since 1982. In August 2016, SST was up to 5° C higher than the 1982-2010 average in regions of the Barents and Chukchi seas and off the east and west coasts of Greenland.

Increasing ocean primary production (conversion of CO_2 to organic material) is also being observed as summer sea ice extent declines, related to enhanced light availability. In 2016, there were widespread positive primary production anomalies throughout the Arctic Ocean and adjacent ice-affected seas, from 5% to 19% above the

2003-2015 average in Hudson Bay and the Barents Sea, respectively. The only negative anomalies were observed in the western (North American) Arctic (-11.9%) and the Sea of Okhotsk (-1.8%). For the period 2003-2015 there are statistically significant primary production trends in the in the eastern (Eurasian) Arctic, Barents Sea, Greenland Sea, Hudson Bay and North Atlantic; the steepest trends are in the eastern Arctic (37.9% increase) and the Barents Sea (34.8% increase).

The waters of the Arctic Ocean are more prone to ocean acidification (OA) compared to the rest of the global ocean, due their cooler water temperatures and unique physical processes (e.g. the formation and melting of sea ice). Even small amounts of human-derived carbon dioxide (CO_2) can cause significant chemical changes that other areas do not experience. Current data indicates that certain areas of the Arctic shelves presently experience prolonged ocean acidification events in shallow bottom waters. These waters are eventually transported off the shelf. As a result, corrosive conditions have been expanding deeper into the Arctic Basin over the last several decades. The inherently short Arctic food web linkages generate an increased urgency in the need to understand the impacts of OA on the Arctic marine ecosystem.

Ice on land, as represented by the Greenland Ice Sheet, saw a continuation of the overall increasing melting trend in 2016, with enhanced melt occurring in the southwest and northeast regions. The onset of surface melt ranked 2nd (after only 2012) over the 37-year period of satellite record (1979 - 2016). The duration of the melt season lasted 30-40 days longer than usual in the northeast and 15-20 days longer along the west coast, compared to the 1981-2010 average. Consistent with the spatial distribution of melt anomalies, the largest area of relatively low albedo (a measure of surface reflectivity) was located along the southwest coast, reaching down to ~20% below the 2000-2009 average.

The spring snow cover extent (SCE) has undergone significant reductions over the period of satellite observations (which start in 1967), particularly since 2005. In 2016, new record low April and May SCE was reached for the North American Arctic. The June SCE was the 3rd lowest on record over both the North American and Eurasian sectors of the Arctic. Warming Arctic surface air temperatures have a clear influence on the timing of snow melt. However, there is also evidence of decreasing pre-melt snow mass (reflective of shallower snow) which may pre-condition the snowpack for earlier and more rapid melt in the springtime.

Satellite observations of tundra greenness (a measure of vegetation productivity and strongly correlated with above-ground biomass) are available since 1982, with 2015 being the most recent year with a complete data set. Long-term trends over this period show greening on the North Slope of Alaska, in the southern Canadian tundra, and in much of the central and eastern Siberian tundra. A decreasing trend in greenness, or "browning", is observed in western Alaska (Yukon-Kuskokwim Delta), the higher-Arctic Canadian Archipelago, and western Siberian tundra.

Warming air temperatures in the Arctic are causing normally frozen ground (permafrost) to thaw. The permafrost is carbon rich and, when it thaws, is a source of the greenhouse gases carbon dioxide and methane. Northern permafrost zone soils contain 1330-1580 billion tons organic carbon, about twice as much as currently contained in the atmosphere. Tundra ecosystems are taking up increasingly more carbon during the growing season over the past several decades, but this has been offset by increasing carbon loss during the winter. Overall, tundra appears to be releasing net carbon to the atmosphere.

Changes in the tundra environment also affect the fauna they support, causing species to shift their distributions, adapt to novel conditions, or die out. Because of this linkage, small mammals, such as shrews, and their parasites can serve as key indicators for anticipating the consequences of environmental variability and changing interactions among species. Some Arctic shrews have recently acquired new parasites indicating poleward shifts of sub-Arctic faunas, and demonstrating increases in both Arctic biodiversity and overall complexity within these novel species networks.

In summary, there are numerous and diverse signals indicating that the Arctic environmental system continues to be influenced by long-term upward trends in air temperature, modulated by natural variability in regional and seasonal anomalies (**Fig. 0.1**). The acceleration of many of these signals, the interdependency of the physical and biological elements of the Arctic system, and the growing recognition that the Arctic is an integral part of the larger Earth system are increasing the pressure for more effective and timely communication of these scientific observations to diverse user audiences. A key to meeting this challenging goal is to more directly convey the syntheses of observations across disciplinary boundaries, to better highlight Arctic system change.

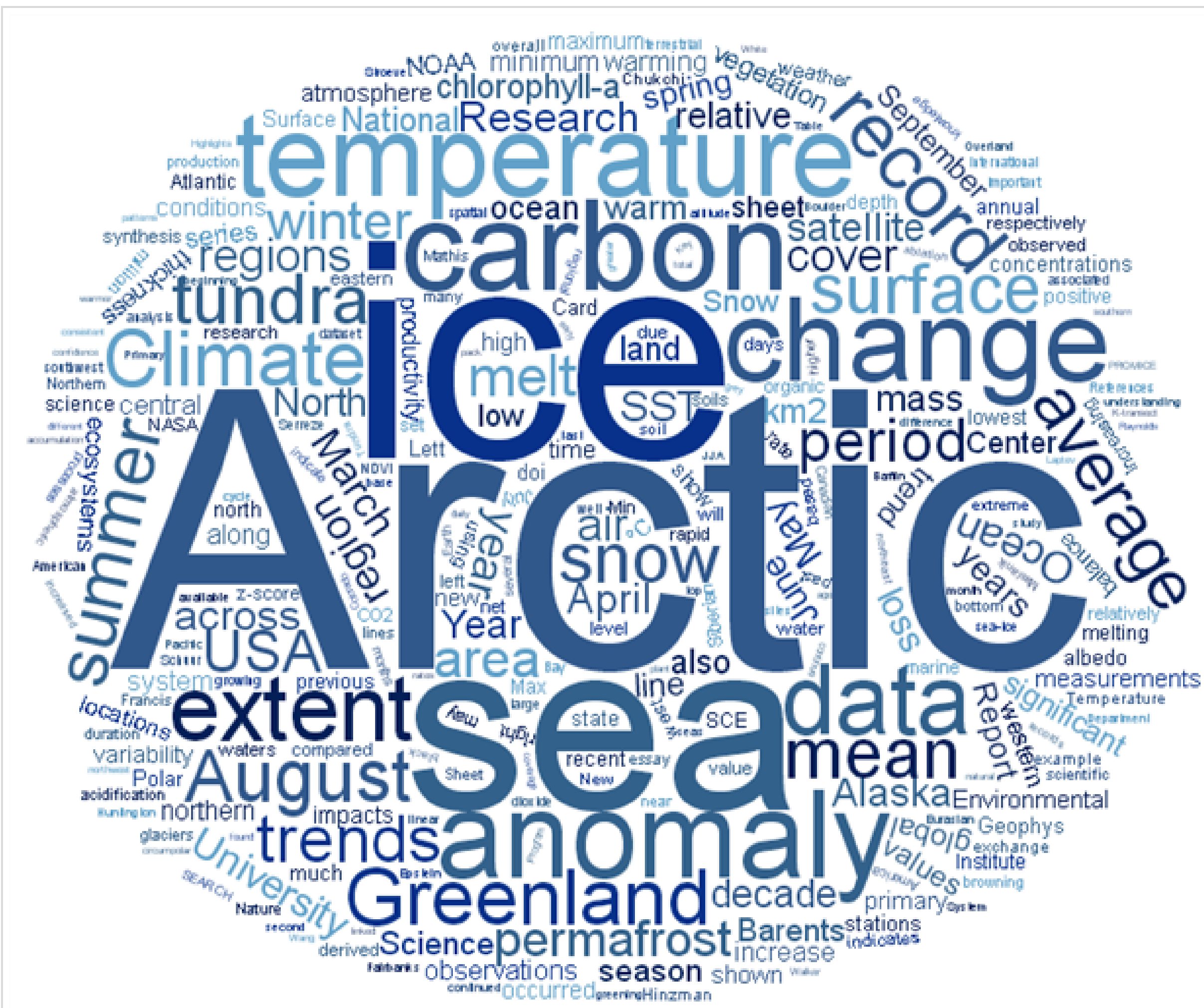

Fig. 0.1. Wordle of 2016 Arctic Report Card. Created with http://www.wordclouds.com/.

Acknowledgments

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December 6, 2016

2016 Addendum

Arctic Report Card 2016 Addendum: October-November 2016

The tight publication schedule of the Arctic Report Card, which includes a peer-review of its content, means the updated observations extend through September of the reporting year. Since some Arctic conditions in Fall 2016 are exceptional, we've added this update to extend the scope through November for those 'Vital Signs' essays where data are available: surface air temperature, surface ocean temperature, sea ice, terrestrial snow cover, and Greenland ice sheet.

Note: This addendum has not been peer-reviewed.

Summary

It is apparent that the record-breaking delay in the freeze up of the sea ice cover in Fall 2016 is associated with unprecedented warm air and ocean surface temperatures. Further, the pattern of snow extent anomalies over land in Fall 2016 is strongly related to the pattern of surface air temperatures.

While October and November established new record lows for Arctic sea ice extent, the exceptionally warm temperature anomalies, which delayed the onset of ice formation, did not extend over Arctic and subarctic land areas. Instead, colder than normal surface temperatures, particularly over Eurasia, allowed terrestrial snow cover extent to be greater than normal. Though relatively warm, surface temperatures over the Greenland ice sheet remained below the melting point and no surface melting was detected during the fall season. Differences in Arctic and subarctic/mid-latitude surface temperature anomalies observing in Fall 2016 were driven by a large scale atmospheric wind pattern that was unusual but not unexpected.

Surface Air Temperature (prepared by J. Overland, NOAA)

The average October-November 2016 air temperature in the lower atmosphere in the central Arctic was 6° C (11° F) above the 1981-2010 climate average, with the highest values spreading into Northwestern Canada. (**Fig. A1**). Daily weather records were exceeded all over the Arctic especially near the Kara Sea, Svalbard, and northern Canada, which were up to 14° C (25° F) higher than normal. It is worth noting that the October-November composite temperature anomaly map, shown is **Fig. A1**, features a pattern that persisted over these months.

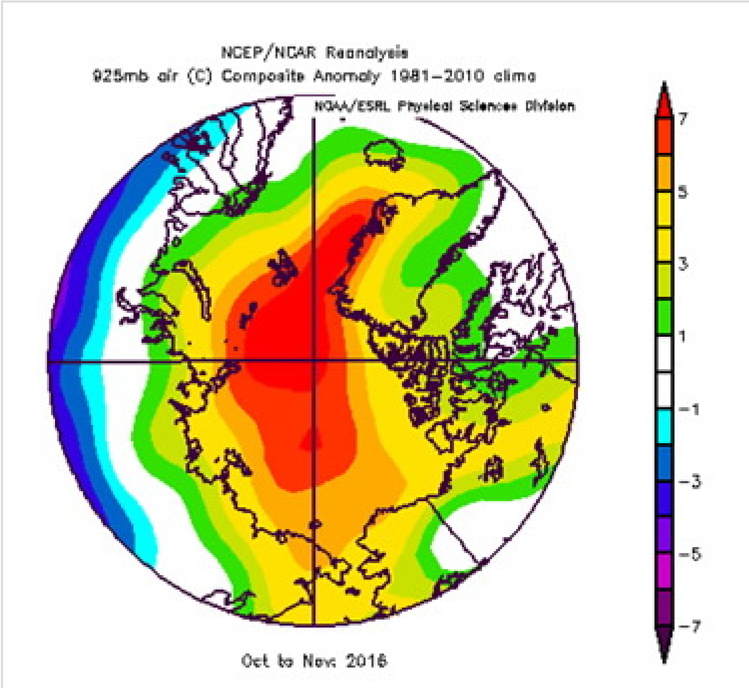

Fig. A1. Lower atmospheric air temperature anomalies for October-November 2016. Consistent with the figures in the *Surface Air Temperature* essay in the full report card, temperatures are from slightly above the surface layer (at 925 mb level) to emphasize large spatial patterns rather than local features. Data are from NOAA/ESRL, Boulder, CO, at http://www.esrl.noaa.gov/psd/.

The immediate cause of the warm Fall 2016 Arctic temperatures were winds bringing in warmer temperatures from mid-latitudes, similar to conditions discussed for the warm temperature extremes from Winter 2016. In Fall 2016 there were deep low pressure systems in the western Bering Sea, as part of the Aleutian low, and from west of Iceland extending into northern Baffin Bay (**Fig. A2**). On the east side of these pressure centers strong winds brought warm conditions up through Bering Strait and across the northern North Atlantic and on to the vicinity of the North Pole.

Conditions in Fall 2016 are similar to conditions in Winter 2016 (see essay on *Surface Air Temperature*). Both Winter and Fall 2016 represent similar extreme temperature conditions tied to an increased interaction between mid-latitude weather and new changes in the Arctic. For Winter 2016 we speculated that this was primarily a random weather event of the tropospheric polar vortex, but the repeat of the pattern in Fall 2016 suggests more complicated interactions.

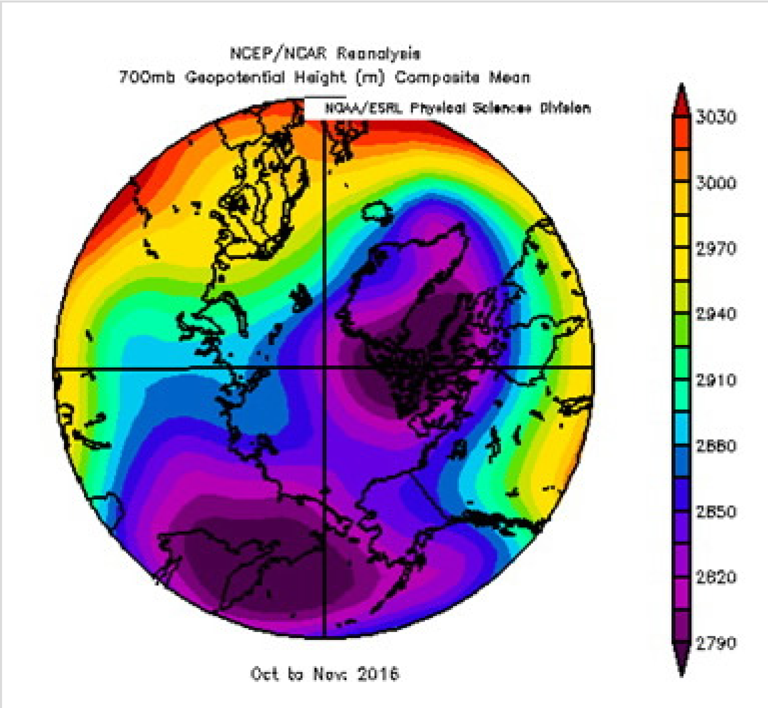

Fig. A2. A split tropospheric polar vortex wind pattern (winds follow contours) for Fall 2016. Winds bring relatively warm air from mid-latitudes across Bering Strait and from the northern North Atlantic. Pattern is similar to that discussed for Winter 2016 (see essay on *Surface Air Temperature*). Data are from NOAA/ESRL, Boulder, CO, at http://www.esrl.noaa.gov/psd/.

Surface Ocean Temperature (M.-L. Timmermans, Yale University)

Mean sea surface temperatures (SST) in September 2016 in ice-free regions ranged from ~0° C just south of the marginal ice zone to around +7° C in the southern Chukchi Sea, and up to +10° C in the Barents Sea (**Fig. A3a**). A similar pattern, although about 1° C cooler in general, was observed for October 2016 mean SST (**Fig. A3b**).

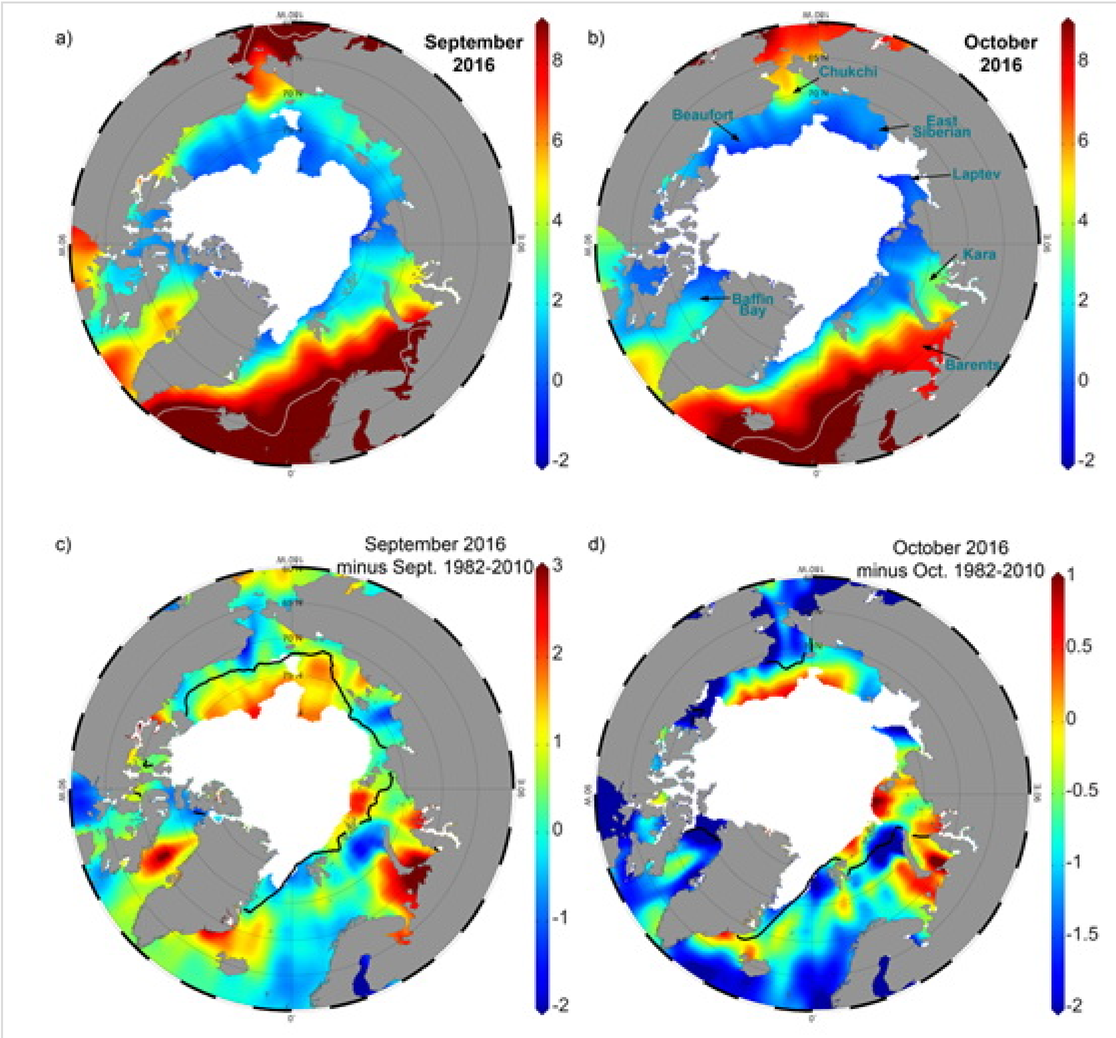

Fig. A3. Mean sea surface temperature (SST, °C) in (a) September and (b) October 2016. White shading is the mean sea ice extent for the respective month, and gray contours indicate the 10° C SST isotherm. SST anomalies (°C) in (c) September 2016 relative to the September mean for the period 1982-2010 and (d) similarly for October 2016. White shading is the 2016 mean ice extent for the respective month and the black contour indicates the median ice edge in the respective month for the period 1982-2010. SST data are from the NOAA Optimum Interpolation (OI) SST Version 2 product (OISSTv2). Sea-ice extent and ice-edge data are from NSIDC.

Anomalously warm SSTs in September 2016 compared to the 1982-2010 September mean (**Fig. A3c**) were largely confined to regions that were ice-covered in the long-term mean, relative to ice-free regions in September 2016. In these regions near the marginal ice zone (i.e. near the edge of the sea ice cover), values were up to +2° C warmer in September 2016 compared to the September 1982-2010 mean. In addition to these regions, SSTs in the Barents Sea were up to +2° C warmer in September 2016 compared to the September 1982-2010 mean, and also warmer in Baffin Bay, and off the east coast of Greenland (**Fig. A3c**).

A similar pattern persists for October 2016 SST anomalies compared to the 1982-2010 October mean (**Fig. A3d**) with anomalously warm temperatures generally just to the south of the October 2016 ice edge; regions that were ice covered in the October 1982-2010 mean. In addition to these regions, SSTs in the Barents and Kara seas were up to +1° C warmer in October 2016 compared to the October 1982-2010 mean (**Fig. A3d**).

Sea Ice (D. Perovich, CRREL)

After an early rapid freeze-up in late September, there was a remarkably slow increase in ice extent during the Fall 2016 (**Fig. A4**). From mid-October to late-November, the ice extent has been the lowest observed since the beginning of the satellite record in 1979. The October 2016 ice extent was 2.55 million km^2 (28%) less than the long term October value, averaged from 1981 to 2010. In November 2016, the ice extent was 1.95 million km^2 (18%) below the 1981 to 2010 average. The ice extent remained well below average into November in the Bering Strait, Chukchi, Barents and Kara Seas, and along Greenland's eastern coast. Results from ESA's CryoSat indicate a low ice volume for November 2016, as well. The low volume is a direct consequence of the reduced ice extent, as the ice that is present is of greater than average thickness (**Fig. A5**).

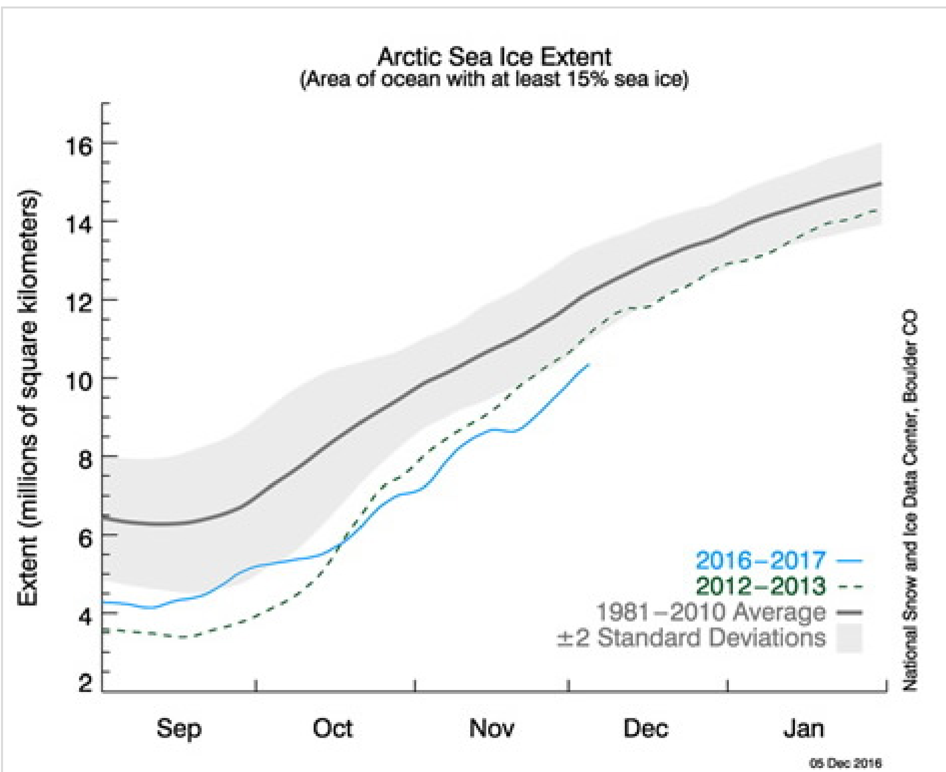

Fig. A4. Arctic sea ice extent during fall freeze-up, comparing 2016 to 2012 and the 1981-2010 average. The graph is from the National Snow and Ice Data Center, Boulder CO.

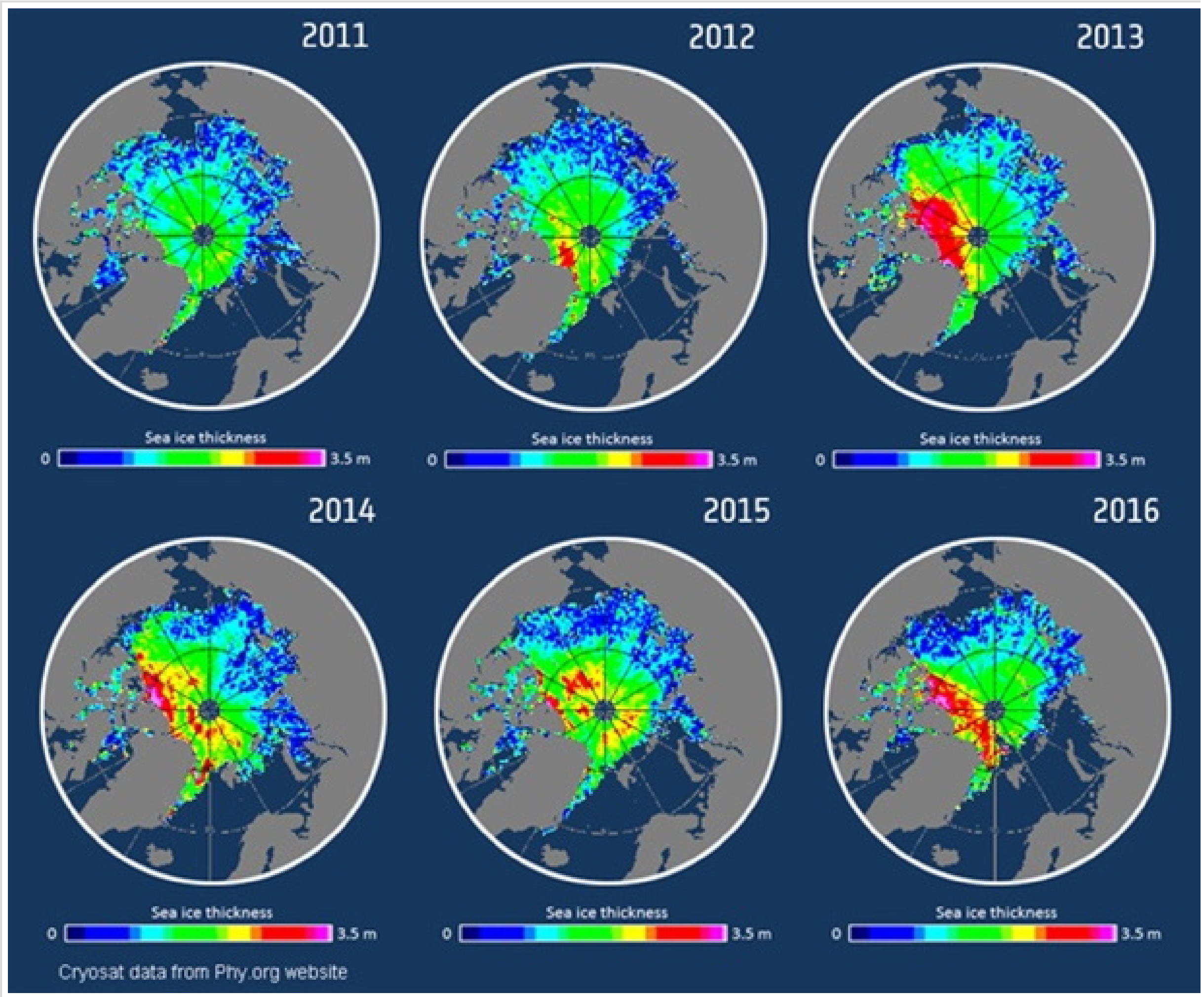

Fig. A5. November Arctic sea ice thickness for the years 2011 to 2016, derived from the European Space Agency's CryoSat. Maps are from the Phys.org website, with data credited to CPOM/ESA.

Terrestrial Snow Cover (C. Derksen, ECCC)

The warm temperature anomalies over the central Arctic in October and November 2016 (**Fig. 1A**), associated with the new record lows for Arctic sea ice extent (**Fig. A4**), did not extend over Arctic and subarctic land areas. Instead, October 2016 surface temperatures over Eurasian Arctic land areas were near normal, a zone of well below average temperatures extended across the Eurasian subarctic and mid-latitude regions, and a region of colder than normal temperatures was over northwestern Canada (**Fig A6a**). This resulted in more extensive than normal snow cover extent over Eurasia and Canada (**Fig. A6c**). Colder than normal surface temperatures continued to extend over much of Eurasia in November 2016 (**Fig. A6b**), which allowed the above normal snow extent established in October to persist in this region (**Fig. A6d**). Air temperatures anomalies over North America were positive in November (**Fig. A6c**), so the above average snow extent observed in October transitioned to largely negative anomalies (**Fig. A6d**). Time series of October and November northern hemisphere snow extent from the NOAA snow chart climate data record showed October and November 2016 to have the third and fourth highest monthly values respectively, in the record (which dates back to 1967).

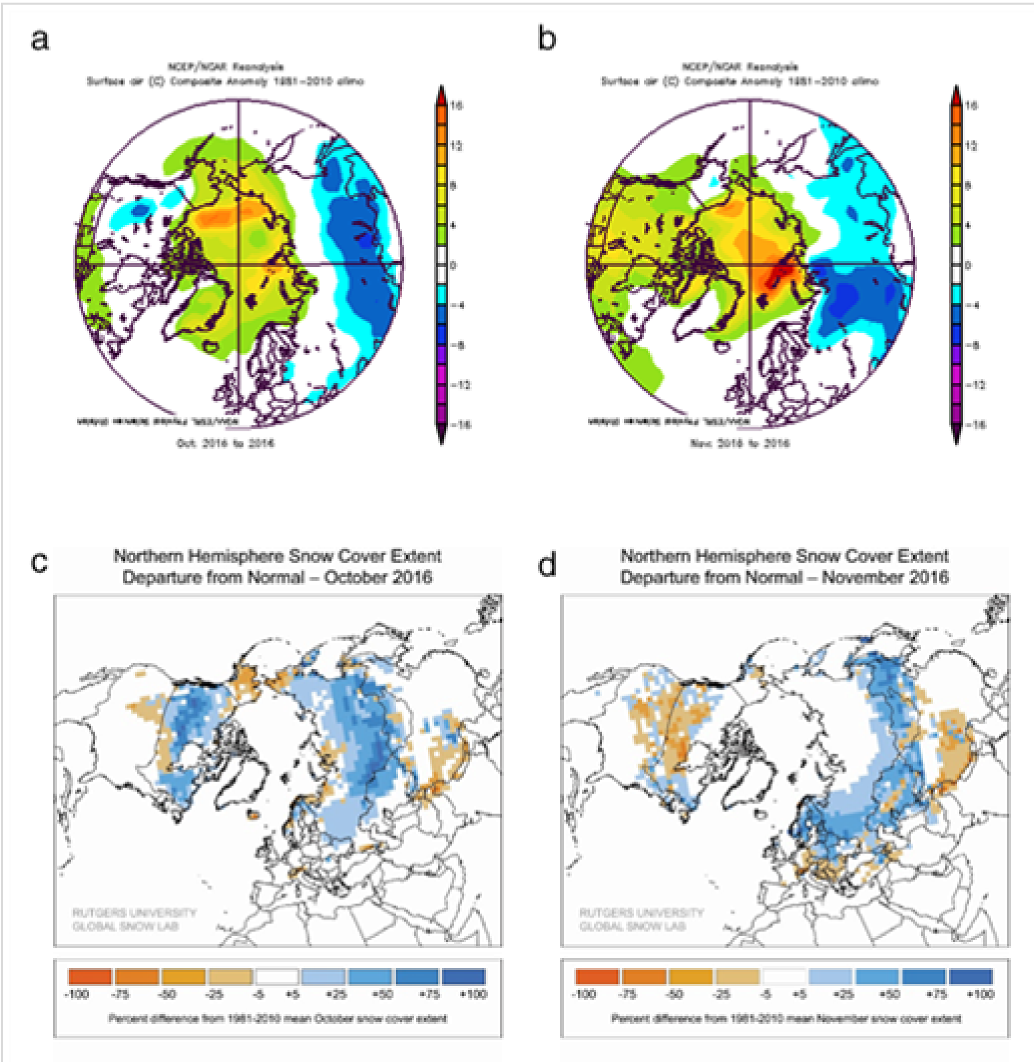

Fig. A6. Monthly surface temperature anomalies for (a) October and (b) November derived from NCEP/NCAR reanalysis (http://www.esrl.noaa.gov/psd/data/reanalysis/reanalysis.shtml). Departures in (c) October and (d) November snow coverage (percent difference from 1981-2010 mean) derived from the NOAA snow chart climate data record (v01r01) (http://climate.rutgers.edu/snowcover/).

Greenland Ice Sheet (M. Tedesco, LDEO)

Air temperature anomalies over the Greenland ice sheet during Fall 2016 were above the 1981-2010 mean by up to +5-6° C in the northeast and up to +1-3° C in the southern portion of the ice sheet (**Fig. A1**). However, these surface temperatures remained below the melting point and no surface melting was detected during fall.

No update for the GRACE data set is available, and estimates of total mass changes during Fall 2016 will be available in April 2017.

December 9, 2016

Surface Air Temperature

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Highlights

- The average annual surface air temperature anomaly over land north of 60° N for October 2015-September 2016 (+2.0° C, relative to a 1981-2010 baseline) was by far highest in the observational record beginning in 1900; this represents a 3.5° C increase since the beginning of the 20th Century.
- Arctic temperatures continue to increase at double the rate of the global temperature increase.
- Extreme Arctic-wide air warm temperatures in winter 2016 (Jan-Mar) greatly exceeding the previous record; several locations showed January anomalies exceeding +8° C. These conditions were primarily due to southerly winds moving warm air into the Arctic from mid-latitudes and the presences of sea ice free areas to the northeast of Novaya Zemlya.
- Neutral to cold temperature anomalies occurred across the central Arctic Ocean in summer 2016; a condition which did not support rapid summer sea ice loss.

Arctic air temperature is an indicator of regional and global climate change. Although there are year-to-year and regional differences in air temperatures due to natural variability, the magnitude and Arctic-wide character of the long-term temperature increase is a major indicator of global warming and the influence of increases in Greenhouse gases (Overland, 2009; Notz and Stroeve, 2016). Here we report on the spatial and temporal variability of Arctic air temperatures during the period October 2015 through September 2016, the 12-month period since the end of the previous Arctic Report Card reporting period.

Mean Annual Land Surface Temperature

The mean annual surface air temperature anomaly (+2.0° C relative to the 1981-2010 mean value) for October 2015-September 2016 for land stations north of 60° N is the highest value in the record starting in 1900 (**Fig.**

1.1). This is an increase of 3.5° C since the beginning of the 20th Century, and the largest annual increase since 1995. Currently, the Arctic is warming at more than twice the rate of lower latitudes (**Fig. 1.1**).

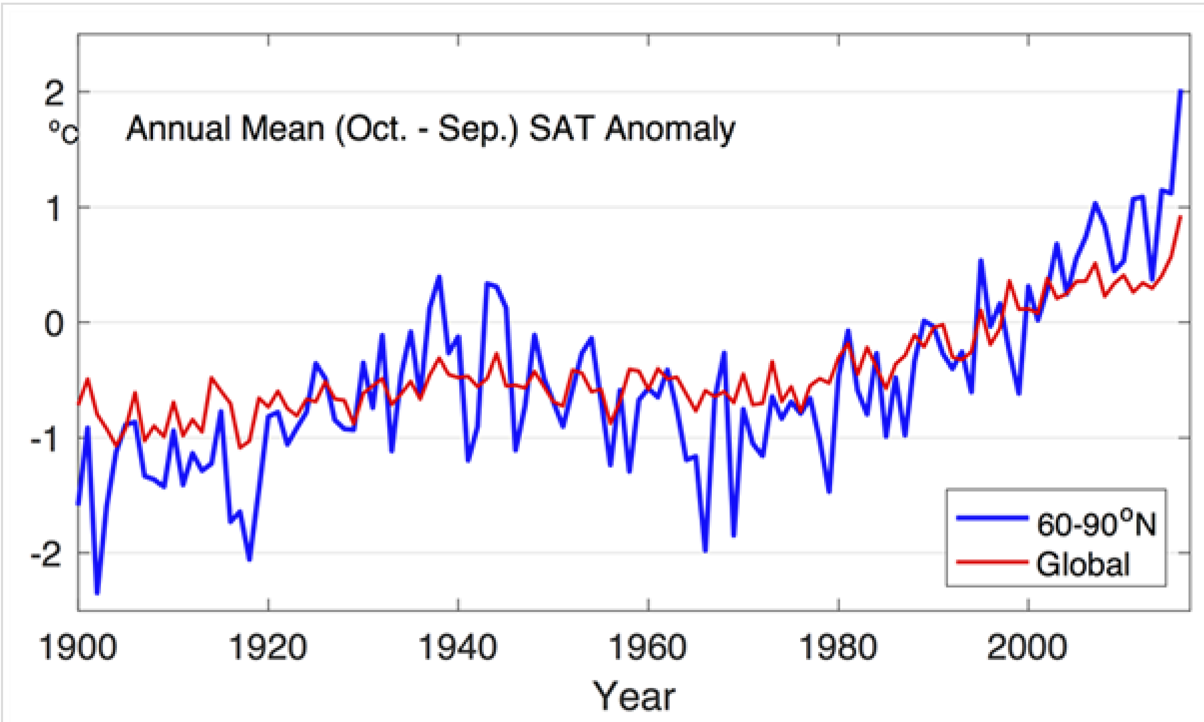

Fig. 1.1. Arctic (land stations north of 60° N) and global mean annual land surface air temperature (SAT) anomalies (in °C) for the period 1900-2016 relative to the 1981-2010 mean value. Note that there were few stations in the Arctic, particularly in northern Canada, before 1940. The data are from the CRUTEM4 dataset, which is available at www.cru.uea.ac.uk/cru/data/temperature/.

The greater rate of Arctic temperature increase than the global increase is referred to as Arctic Amplification. Mechanisms for Arctic Amplification include: reduced summer albedo due to sea ice and snow cover loss, the increase of total water vapor content in the Arctic atmosphere, a decrease of total cloudiness in summer and increase in winter (Makshtas et al. 2011), the additional heat generated by newly sea-ice free ocean areas that are maintained later into the autumn (Serreze and Barry 2011), and the lower rate of heat loss to space in the Arctic relative to the sub-tropics due to lower mean temperatures (Pithan and Mauritsen 2014). More research is needed to quantify such contributions. Further, Arctic warming has been influenced by past air pollution reductions in Europe (Acosta Navarro et al. 2016). However, increases in 2016 Arctic temperatures has a strong natural variability component from the supply of warm air originating in mid-latitudes, as described in the next two sections.

Seasonal Surface Air Temperature Variation

Seasonal air temperature variations are divided into autumn 2015 (October, November, December [OND]), and winter (January, February, March [JFM]), spring (April, May, June [AMJ]) and summer (July, August, September

[JAS]) of 2016. These seasonal divisions are chosen for SAT to coincide with some key Arctic climate (December is still cooling down) and sea ice divisions (September is summer sea ice minimum month). Autumn, Spring and especially Winter show extensive positive temperature anomalies across the central Arctic (**Fig. 1.2**).

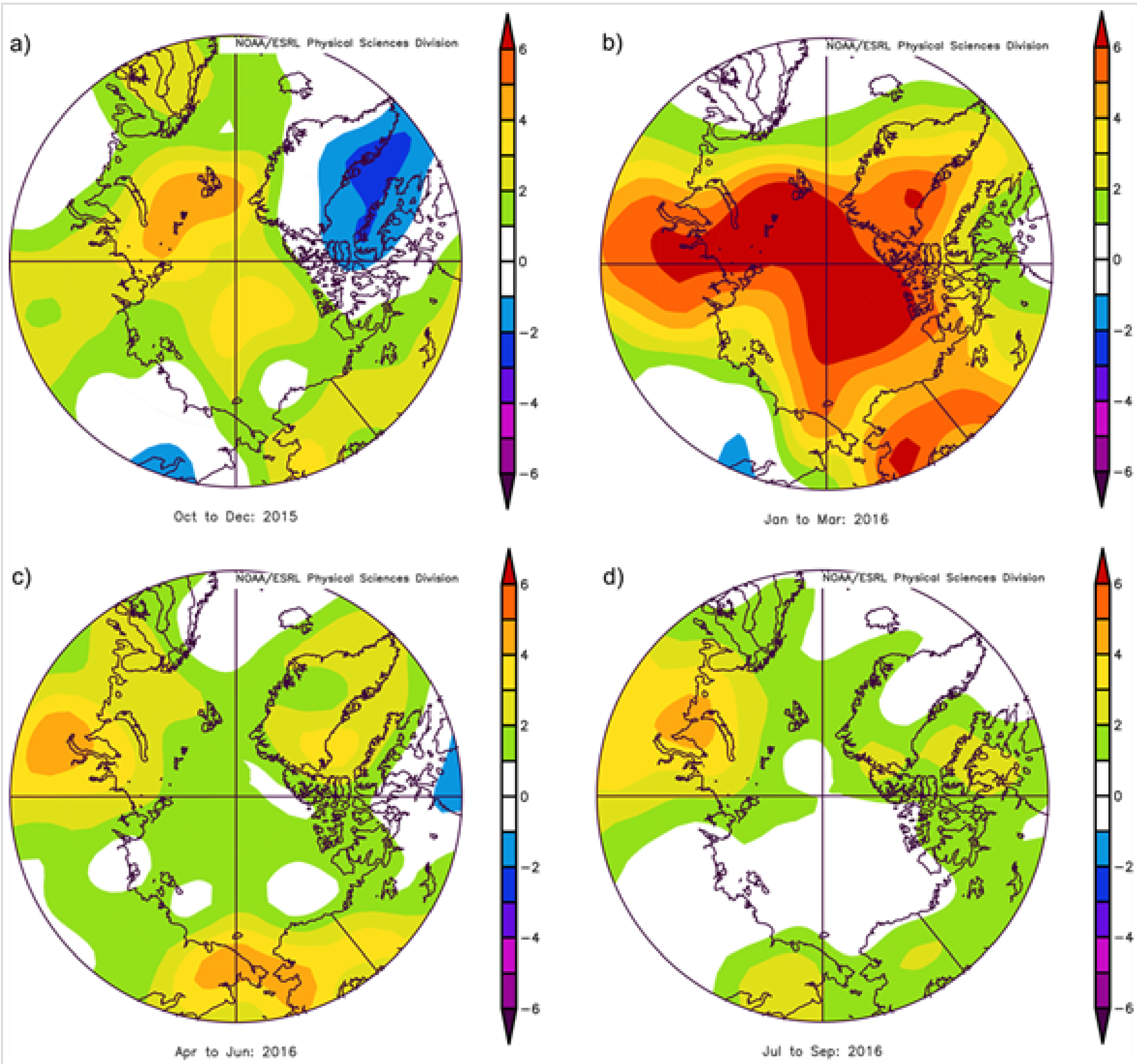

Fig. 1.2. Seasonal anomaly patterns for near surface air temperatures (in °C) relative to the baseline period 1981-2010 in (a, top left) autumn 2015, (b, top right) winter 2016, (c, bottom left) spring 2016 and (d, bottom right) summer 2016. Temperatures are from slightly above the surface layer (at 925 mb level) to emphasize large spatial patterns rather than local features. Data for this and the following figures are from NOAA/ESRL, Boulder, CO, at http://www.esrl.noaa.gov/psd/.

Autumn 2015 (OND). A broad swath of warm temperature anomalies stretched across the Arctic, extending from the Pacific to the Atlantic sectors (**Fig. 1.2a**). The warmest temperature anomalies, centered on western Alaska and the Atlantic sector in northern Barents Sea, were due to advection of warm air from the south. In the Spitsbergen area the 3-month mean temperatures are 4 to 6° C above 1961-90 average.

Winter 2016 (JFM). Extreme record warm temperature anomalies extend across the Arctic, from the Pacific to the Atlantic sector (**Fig. 1.2b**). For January, the Arctic-wide averaged temperature anomaly was 2.0° C above the previous record of 3.0° C, and local extremes exceeded 8° C (Overland and Wang 2016). The warmest temperature anomalies were centered on Alaska, Svalbard in the Atlantic sector, and the central Arctic. In the Spitsbergen area the 3-month winter mean temperatures were 8 to 11° C above 1961-90 average. Temperature anomalies for coastal Greenland locations were warmer in the north (see essay on *Greenland Ice Sheet*). In contrast, there was a severe cold temperature surge into China from the north (**Fig. 1.2b**). The only region that did not experience warm winter temperature extremes was Scandinavia, where temperatures were near the mean. The cause of these extreme temperature are described further in the section below on *Arctic and Mid-latitudes Connections*.

Spring 2016 (AMJ). A broad swath of warm, but smaller magnitude, temperature anomalies continued to stretch across the central Arctic. Extremes continued in the periphery near Bering Strait, northern Greenland and eastern Russia (**Fig. 1c**).

Summer 2016 (JAS). The warm winter/spring conditions of 2016 did not continue through the summer months. Summer 2016 also contrasts to warm conditions in much of the previous decade (**Fig. 1.2d**). Neutral to cold anomalies occurred across the central Arctic in summer 2016, which did not support rapid summer sea ice loss (see essay on *Sea Ice*). Subarctic regions had warm anomalies in central Alaska, east Siberia, northern Baffin Bay and the southern Kara Sea. Alaska had the second warmest summer on record with more convective type clouds than normal. Alaska daily temperature anomalies (25-station average) were above normal during January-September 2016. August 2016 was the 69th successive month with temperature above the 1961-1990 climatology at Svalbard Airport.

Arctic and Mid-latitudes Connections

Record winter temperature conditions in the Arctic were primarily a response to southerly winds advecting warm air into the Arctic from mid-latitudes. This pattern began in the autumn 2015 with the development of strong low pressure in the North Atlantic and the Bering Sea that directed air trajectories toward the Arctic. Alaska and the Atlantic Arctic were anomalously warm in fall 2015, and continued to be so into winter and spring 2016 (**Figs. 1.2a, 1.2b** and **1.2c**). The positive near-surface air temperature anomalies were associated with a persistent pattern of geopotential height field characterized by low values in the Barents Sea and Aleutian Islands and higher values in the central Arctic (**Fig. 1.3**). Winds tend to follow the contours of geopotential heights counter-clockwise around low values. Consequently, warm air over the northeast Pacific Ocean was advected into and across Alaska; north of the Kara Sea warm southerly winds extended across the central Arctic past the North Pole (i.e. follow purple contours in **Fig. 1.3**). Normally there is a single large Arctic region of low geopotential heights which establishes a tropospheric polar vortex (see for example Overland et al. 2015). In winter 2016 higher geopotential heights in the central Arctic split the polar vortex into two pieces. The persistence of warm Arctic temperatures was in part caused by a positive feedback, where advection of warm air temperatures contributes to increasing the geopotential heights over the central Arctic, which in turn helped to maintain the unusual 2016 wind pattern. Higher geopotential heights in the sub-Arctic over land along 90 E brought relatively colder air to western China.

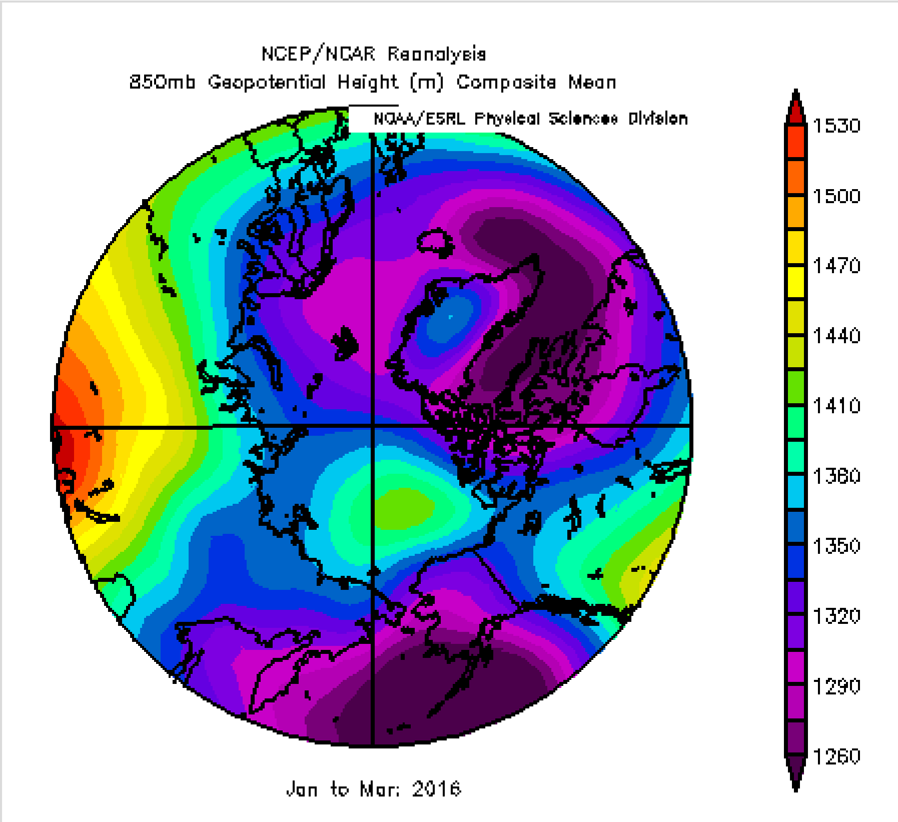

Fig. 1.3. Geopotential height (850 mb) from January-March 2016. The normally continuous tropospheric polar vortex of low heights (purple shading) over the central Arctic was split in two parts in winter 2016, giving rise to southerly winds and extreme record temperatures over the central Arctic.

Weather Connections to Sea Ice and Greenland

Extreme warm winter and spring air temperatures in 2016 contributed to record low sea ice extents at the beginning of the summer melt season (see essay on *Sea Ice*). But continued record sea ice loss over the summer was not to be. Neutral to cold temperature anomalies (**Fig. 1.2d**) contrasted to warmer conditions in previous major sea ice loss years (2007, 2012) (Overland et al 2015). The past decade often showed higher sea level pressures (SLP) and warm summer conditions. Summer Arctic 2016 was dominated by low SLP (**Fig. 1.4**), responsible for two major storms during August. While 2016 is the second lowest summer sea ice extent, the cause contrasted with previous major minimum years; 2016 conditions depended on low initial spring sea ice, rather than supportive summer weather with high SLP, as in recent years.

Summer 2016 saw an increased advection of relatively warm air into west Greenland (**Fig 1.2d**), consistent with having the second-highest Greenland Blocking Index (GBI) on record (beginning in 1851), after only 2012 (i. e. Hanna et al. 2016). The GBI reflects the distribution of high geopotential heights, responsible for diverting the jet stream location. As a result, melting occurred more frequently than average (relative to Julys spanning 1981 to 2010) along the northern half of the ice sheet (see essay on *Greenland Ice Sheet*).

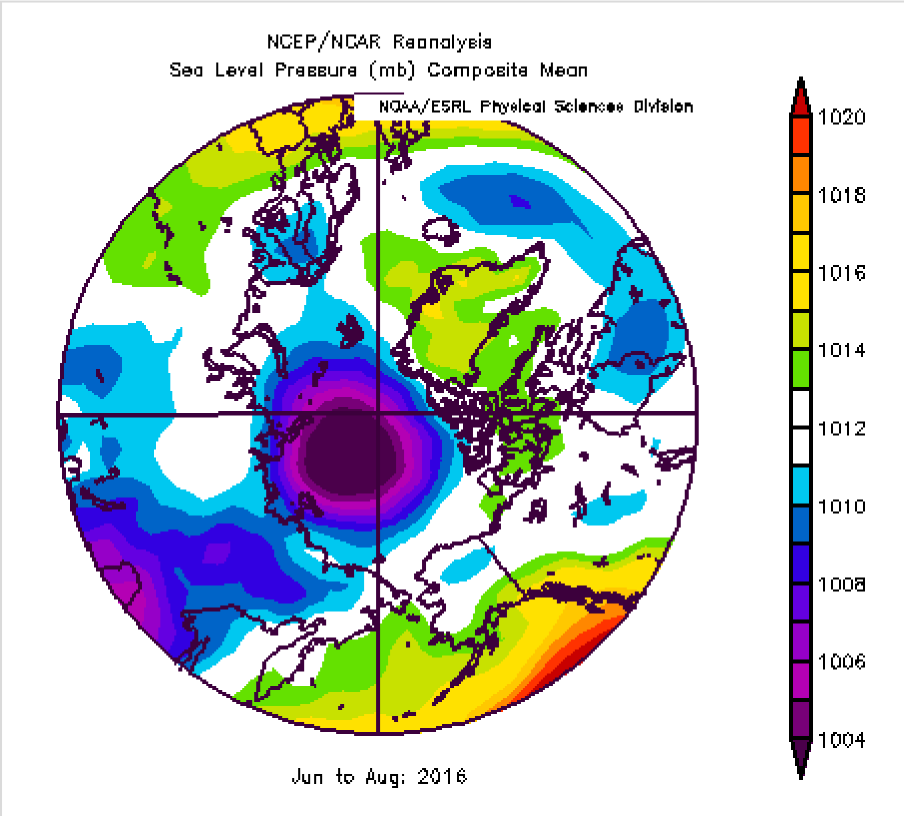

Fig. 1.4. Low sea level pressure dominated the central Arctic during summer 2016 in contrast to most years in the previous decade.

References

Acosta Navarro, J. C. et al. (2016), Amplification of Arctic warming by past air pollution reductions in Europe. *Nature Geosci.* 9, 277-281.

Hanna, E., Cropper, T. E., Hall, R. J. and Cappelen, J. (2016), Greenland Blocking Index 1851-2015: a regional climate change signal. *Int. J. Climatol*., doi: 10.1002/joc.4673.

Makshtas A. P., I. I. Bolshakova, R. M. Gun, O. L. Jukova, N. E. Ivanov, and S. V. Shutilin, 2011: Climate of the Hydrometeorological Observatory Tiksi region. In *Meteorological and Geophysical Investigations*. Paulsen, 2011, 49-74.

Nordli, Ö, R. Przybylak, A. E. J. Ogilvie, and K. Isaksen (2014) Long-term temperature trends and variability on Spitsbergen: the extended Svalbard Airport temperature series, 1898-2012. *Polar Res.*, 33, 21349.

Notz, D., and J. Stroeve, (2016), Observed Arctic sea-ice loss directly follows anthropogenic CO_2 emission. Science, 10.1126/DOI: 10.1126/science.aag2345.

Overland, J. E., 2009: The case for global warming in the Arctic. In *Influence of Climate Change on the Changing Arctic and Sub-Arctic Conditions*, J. C. J. Nihoul and A. G. Kostianoy (eds.), Springer, 13-23.

Overland, J. E., E. Hanna, I. Hanssen-Bauer, B. -M. Kim, S. -J. Kim, J. Walsh, M. Wang, and U. Bhatt, 2015: Air Temperature. In *Arctic Report Card: Update for 2015*, ftp://ftp.oar.noaa.gov/arctic/documents/ArcticReportCard_full_report2014.pdf.

Overland, J. E., and M. Wang (2016): Recent extreme Arctic temperatures are due to a split polar vortex. *J. Climate*, 29(15), 5609-5616, doi: 10.1175/JCLI-D-16-0320.1.

Pithan, F. and T. Mauritsen, 2014: Arctic amplification dominated by temperature feedbacks in contemporary climate models. *Nature Geoscience*, doi: 10.1038/ngeo2071.

Serreze, M., and R. Barry, 2011: Processes and impacts of Arctic amplification: A research synthesis. *Global and Planetary Change*, 77, 85-96.

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Terrestrial Snow Cover

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Highlights

- Spring (April, May, June) snow cover extent (SCE) over Arctic land areas has undergone significant reductions over the period of satellite observations (which start in 1967), particularly since 2005. In 2016, new record low April and May SCE was reached for the North American Arctic, with May SCE falling below 4 million km^2 for the first time in the satellite era; the 3rd lowest June SCE occurred over both the North American and Eurasian sectors of the Arctic.
- Total Arctic SCE in June fell below 3 million km^2 for the sixth time in the past seven years due to earlier than normal snow melt across Alaska, the western Canadian Arctic, and Scandinavia.
- Warming Arctic surface temperatures have a clear influence on the timing of snow melt, although there is also evidence of decreasing pre-melt snow mass (reflective of shallower snow) which may pre-condition the snowpack for earlier and more rapid melt.

Snow cover is a defining characteristic of the Arctic land surface for up to 9 months each year, evolving from complete snow cover in the winter to a near total loss of snow cover by the summer. Highly reflective snow cover acts to cool the climate system, effectively insulates the underlying soil, and stores and redistributes water in solid form through the accumulation season before spring melt. Snow on land in spring has undergone significant reductions in areal extent during the satellite era, which impacts the surface energy budget, ground thermal regime (with associated effects on geochemical cycles), and hydrological processes. The loss of spring snow cover is a clear indicator of change in the terrestrial cryosphere, much in the same way summer sea ice loss is indicative of changes in the marine cryosphere.

Snow cover extent (SCE) anomalies for the Arctic (land areas north of 60N) spring (April, May, June) 2016 were computed separately for the North American and Eurasian sectors of the Arctic (**Fig. 2.1a-c**). The anomalies are assessed relative to the 1981-2010 reference period, from the NOAA snow chart climate data record, which extends from 1967 to present (maintained at Rutgers University; Estilow et al., 2015; http://climate.rutgers.edu/snowcover/). SCE anomalies over the North American sector of the Arctic were strongly negative in all three months: new record low anomalies were set for April and May, with the third lowest values in the NOAA dataset observed in June. Eurasian SCE anomalies were also negative in all three spring months, reaching the third lowest in the NOAA time series in June.

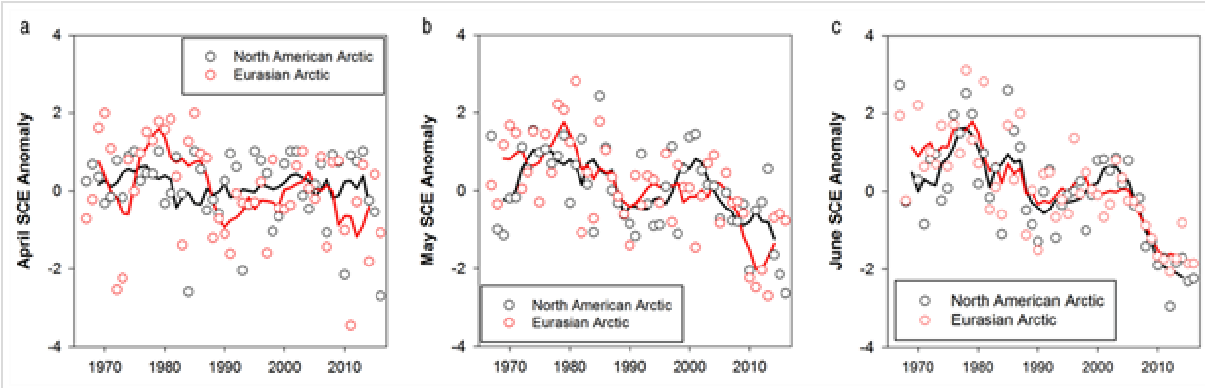

Fig. 2.1. Monthly snow cover extent (SCE) for Arctic land areas (>60° N) from the NOAA snow chart Climate Data Record (CDR) for (a, left) April (b, center) May and (c, right) June from 1967 to 2016. Anomalies are relative to the average for 1981-2010 and standardized (each observation differenced from the mean and divided by the standard deviation and thus unitless). Solid black and red lines depict 5-yr running means for North America and Eurasia, respectively.

Snow cover duration (SCD) departures (**Fig. 2.2**) derived from the NOAA daily Interactive Multisensor Snow and Ice Mapping System (IMS) snow cover product (Helfrich et al., 2007) show earlier snow cover onset in the fall over much of the Arctic (defined as land areas north of 60° N), except Scandinavia. Spring snow cover duration departures were primarily negative (earlier snow melt), with the strongest departures over Alaska, the western Canadian Arctic, and Scandinavia/western Russia. This is consistent with the pattern of positive surface temperature anomalies in spring, which were positive over all Arctic land areas with the exception of eastern North America (see essay on *Surface Air Temperature*).

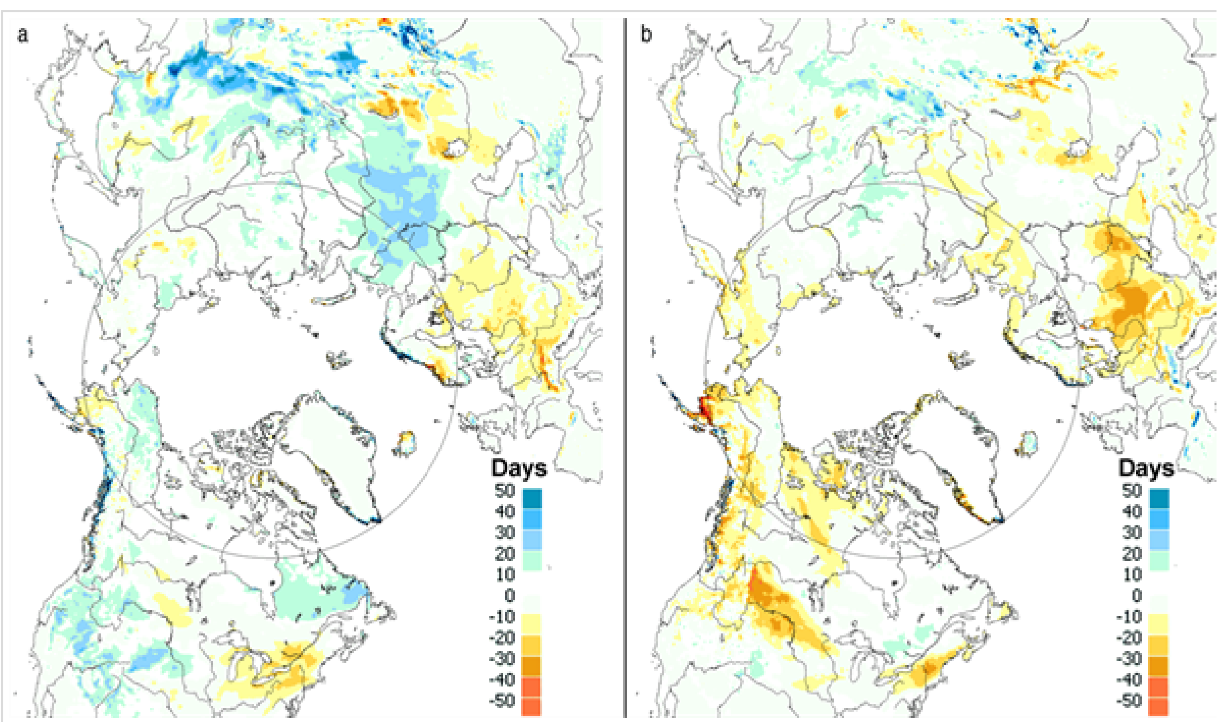

Fig. 2.2. Snow cover duration (SCD in days) departures (difference from 1998-2010 mean) from the NOAA IMS data record fo the 2015-2016 snow year: (a, left) fall; and (b, right) spring. The grey circle marks the latitude 60° N; land north of this defines Arctic land areas considered in this study.

Snow depth anomalies (**Fig. 2.3**; derived from the Canadian Meteorological Centre daily gridded global snow depth analysis; Brasnett, 1999) show a pattern similar to that observed in 2015 (Derksen et al., 2016): negative anomalies (i.e. relatively low snow depth) across the sub-Arctic surrounded mainly positive anomalies over the high latitude regions of Siberia and North America in March and April. By May, the North America snow depth anomalies changed to strongly negative (mean anomaly of -10.7%) consistent with the record low SCE values reported above. May snow depth anomalies over Eurasia were near normal (-0.6%) but plummeted to -29.2% in June.

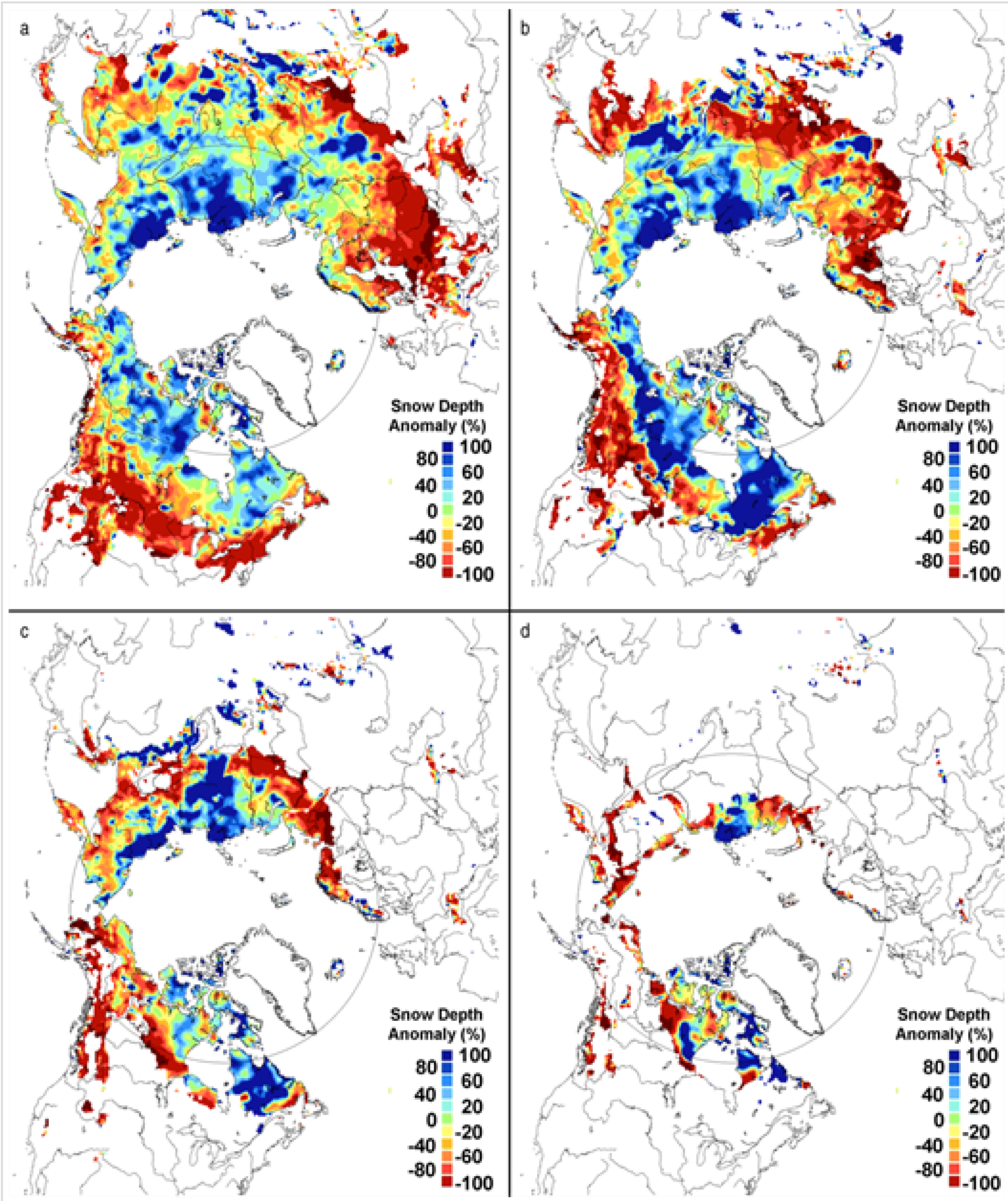

Fig. 2.3. Snow depth anomaly (% of the 1999-2010 average) in 2016 from the Canadian Meteorological Centre (CMC) snow depth analysis for (a, top left) March, (b, top right) April, (c, bottom left) May, and (d, bottom right) June. The grey circle marks the latitude 60° N; land north of this defines Arctic land areas considered in this study.

Time series of pan-Arctic snow extent from the NOAA dataset for May and June are shown in **Fig. 2.4a** over the 1979-2016 time period, which corresponds to the satellite passive microwave data record used to monitor sea ice variability and change. Although falling below 11 million km^2 only three times between 1979 and 2009, May SCE has been below this level every year since 2009. Until 2008, June snow cover was below 4 million km^2 only once (1990), yet it has been below this value every year since. The rate of change in May SCE in the NOAA snow chart data record is now -5.0% per decade, which is significant (95%) but dwarfed by the rate of -17.8%

per decade in June, which exceeds the pace of summer sea ice reductions in September (-13.3% per decade) (see essay on *Sea Ice*). There is currently no evidence of any physical links between changes in snow and sea ice extent but it is instructive to compare the rates of observed change over the terrestrial (snow) and marine (sea ice) components of the cryosphere.

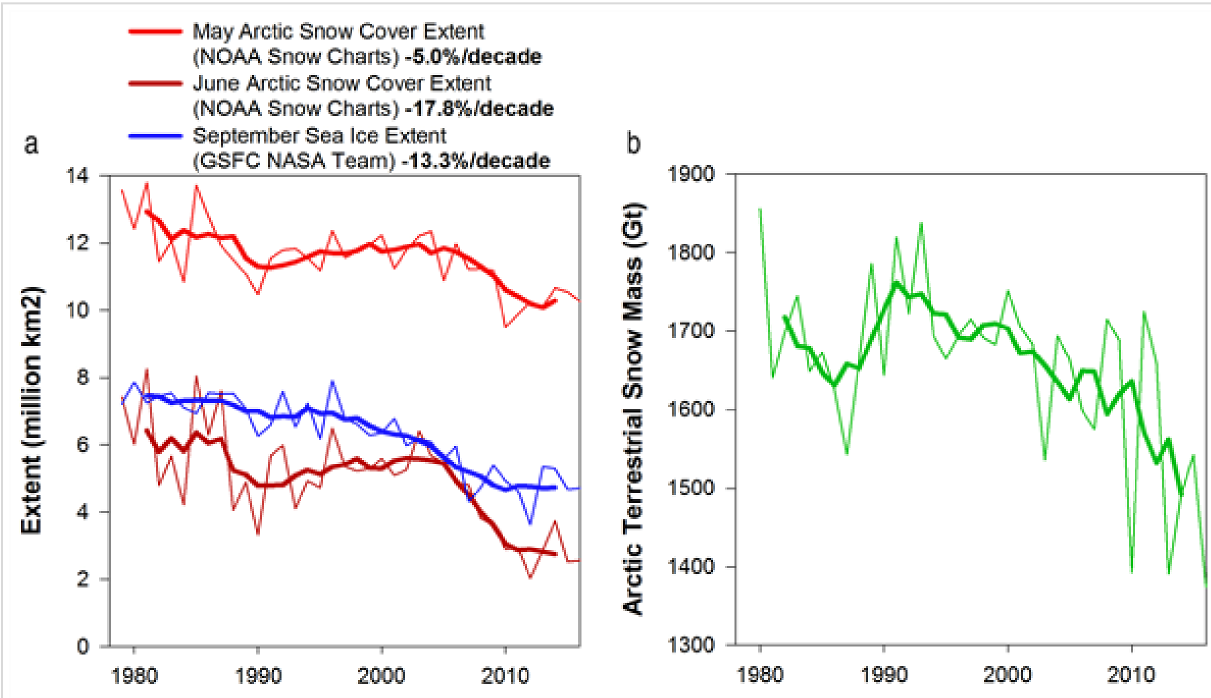

Fig. 2.4. (a) Arctic snow cover extent for May and June, and September Arctic sea ice extent. Sea ice extent for 1979-2015 a from the NASA Team algorithm (Cavalieri et al, 1996); ice extent estimates for 2016 are produced from real time data (Maslanil and Stroeve, 1999). (b) Arctic terrestrial snow mass time series for April, 1980-2016, from the GlobSnow data record (Takala al., 2011). Bold lines are 5-year running means for all datasets.

The link between temperature and snow cover extent in spring is very logical and straightforward: there is a strong association between the magnitude of spring warming and snow cover extent loss in both observational datasets and climate model simulations (Thackeray et al., in press). There is also evidence of decreasing pre-melt snow mass (less water stored in solid form by snow) in the GlobSnow data record (Takala et al., 2011) which combines surface snow depth observations from weather stations with satellite passive microwave measurements. The trend in April snow mass (the month of peak pre-melt Arctic snow mass) is -4.3% per decade, with April 2016 having the lowest value in the record (**Fig. 2.4b**). While early snow melt in previous years occurred despite above average snow mass (for example, 2011 and 2012) a shallower snowpack combined with above average temperatures does create ideal conditions for early and rapid snow melt, reflected in the new record low SCE values observed in 2016.

Seasonal snow cover in the Arctic is in a period of rapid change. We rely heavily on satellite observations and models to monitor and understand snow processes, variability, and trends across Arctic land areas because of the very sparse network of surface observations. It remains a challenge, particularly in an era of decreasing conventional observations of Arctic snow, to disentangle the complex interplay between snow extent, snow mass, surface temperature, and snowfall anomalies which have such an important impact on Arctic land surface processes.

References

Brasnett, B., 1999: A global analysis of snow depth for numerical weather prediction. *J. Appl. Meteorol.*, 38, 726-740.

Cavalieri, D. J., C. L. Parkinson, P. Gloersen, and H. J. Zwally, 1996, updated yearly: Sea Ice Concentrations from Nimbus-7 SMMR and DMSP SSM/I-SSMIS Passive Microwave Data, Version 1 [1979-2014]. Boulder, Colorado USA. NASA National Snow and Ice Data Center Distributed Active Archive Center. http://dx.doi.org/10.5067/8GQ8LZQVL0VL.

Derksen, C., R. Brown, L. Mudryk, and K. Luojus, 2016: Terrestrial Snow (Arctic). In State of the Climate in 2015. *Bull. Am. Met. Soc.*, 97, S145-S147.

Estilow, T. W., A. H. Young, and D. A. Robinson, 2015: A long-term Northern Hemisphere snow cover extent data record for climate studies and monitoring. *Earth Sys. Sci. Data*, 7.1: 137-142.

Helfrich, S., D. McNamara, B. Ramsay, T. Baldwin, and T. Kasheta, 2007: Enhancements to, and forthcoming developments in the Interactive Multisensor Snow and Ice Mapping System (IMS). *Hydrol. Process.*, 21, 1576-1586.

Maslanik, J. and J. Stroeve, 1999, updated daily: Near-Real-Time DMSP SSMIS Daily Polar Gridded Sea Ice Concentrations, Version 1. [September 2015]. Boulder, Colorado USA. NASA National Snow and Ice Data Center Distributed Active Archive Center. http://dx.doi.org/10.5067/U8C09DWVX9LM.

Takala, M., K. Luojus, J. Pulliainen, C. Derksen, J. Lemmetyinen, J. -P. Kärnä, and J. Koskinen, 2011: Estimating northern hemisphere snow water equivalent for climate research through assimilation of space-borne radiometer data and ground-based measurements. *Remote Sens. Environ.*, 115, 3517-3529.

Thackeray, C., C. Fletcher, L. Mudryk, and C. Derksen, In press: Quantifying the uncertainty in historical and future simulations of Northern Hemisphere spring snow cover. *J. Clim.* doi: http://dx.doi.org/10.1175/JCLI-D-16-0341.1.

November 15, 2016

Greenland Ice Sheet

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Highlights

- The 2016 melt season on the Greenland ice sheet continued the overall increasing melting trend, with enhanced melt occurring in the southwest and northeast regions.
- The onset of surface melt ranked 2nd (after 2012) over the 37-year period of record (1979-2016), with the duration of the melt season longer than usual in the northeast (~30-40 days) and along the west coast (~15-20 days). Ice loss measured in-situ at high elevation in the southwest was the 2nd highest in the 27-year measurement period, beginning in 1990.
- Albedo was the 5th lowest in the 17-year period of record (beginning in 2000) for both July and summer averages, with particularly low values along the southwest coast.
- Widespread above average surface air temperatures were observed in the west and the south, with many observational sites setting new records for the spring (MAM) and summer (JJA) seasons and in individual months.

Surface Melting

The Greenland ice sheet plays a crucial role globally and locally, impacting the surface energy budget and climate and weather and contributing to current and future sea level rise. Estimates of the spatial extent of surface melt across the Greenland ice sheet (GrIS), over the period 1979 to 2016, are derived from brightness temperatures measured by the Special Sensor Microwave Imager/Sounder passive microwave radiometer (e.g., Mote 2007, Tedesco et al. 2013). Though 2016 was not a record-breaking year in terms of melt extent and duration, it extends the overall increasing melting trend. The updated trend for melt extent over the period 1979-2016 over the whole Greenland ice sheet is +15,824 km^2/yr.

In 2016, there was an early start to the melt season on April 10th, with melt extent in April reaching values typical of early June. The melt onset date in 2016 is ranked second, by only a few days, to the melt onset day in 2012 on April 4th. Periods of extensive melt (exceeding two standard deviations above the mean) were also recorded in mid-May and in June (**Fig. 3.1a**). The melt extent for the period June through August (JJA) 2016 was above the 1981-2010 average on 66% of days, compared to 54% of the days during the same period in 2015. The anomaly of the number of days when surface melt occurred with respect to the 1981-2010 period reached its peak in the northeast region (**Fig. 3.1b**). The number of melt days was also anomalously high along the west and southwest regions, though not as pronounced as in previous years.

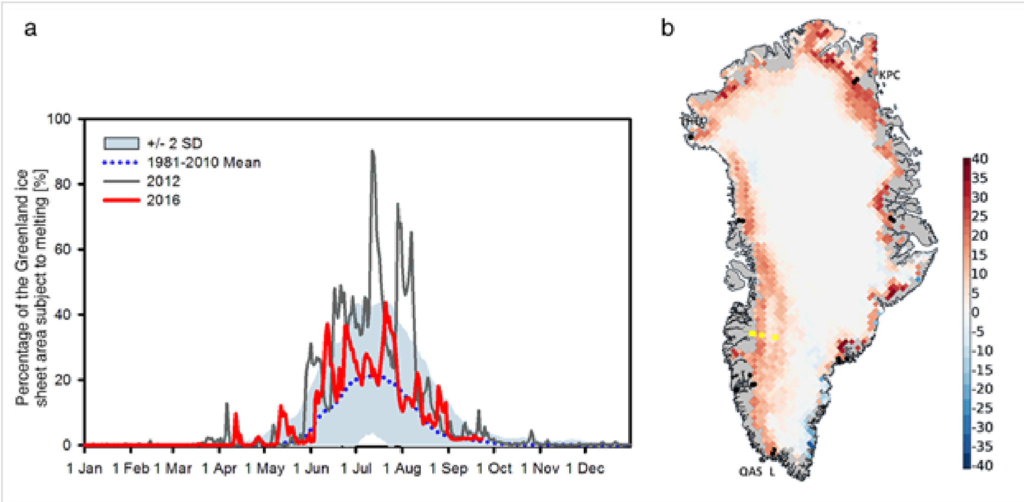

Fig. 3.1. (a) Spatial extent of melt from SSMIS as a percent of the ice sheet area during 2016 (red line) and 2012 (grey line), the 1981-2010 mean spatial extent of melt (dashed blue line), and ± 2 standard deviations of the mean (shaded). (b) Map of the anomalies of the number of melting days obtained from passive microwave data for 2016 with respect to the 1981-2010 mean. Black dots represent the locations of selected PROMICE stations and yellow squares show the location of the K-transect stations. Both plots were produced in conjunction with the National Snow and Ice Data Center.

Surface Mass Balance

The mass balance year 2015-2016 along the K-transect (van de Wal et al. 2005, 2012), located in the southwest part of the GrIS, was characterized by a very high ablation rate at high elevations (**Fig. 3.2a**). In general, there has been an upward migration of the equilibrium line altitude (e.g., the altitude at which the surface mass balance is zero). In 2016, the surface at S9—originally located along the equilibrium line altitude of 1500 m. a.s.l. at the beginning of the time series—showed areas of exposed bare ice, consistent with anomalous melting conditions. Since measurements started in 1991, the ablation rates averaged over the transect have only been higher than 2016 during the 2009-2010 mass balance year, when ablation rates at the edge of the ice sheet were also extraordinary high (**Fig. 3.2b**). The high rates in the upper ablation area can be explained by a combination of limited winter snow accumulation and early melt onset. The latter created a relatively low albedo and, hence, high melt rates. The mass balance gradient along the K-transect (e.g., slope of the SMB plot at different elevations in **Fig. 3.2b**, 3.1 mm water equivalent of ablation per m of height difference per year) in 2016 was comparable to 2015, with the major difference being that in 2016 the equilibrium line altitude was located at a higher elevation.

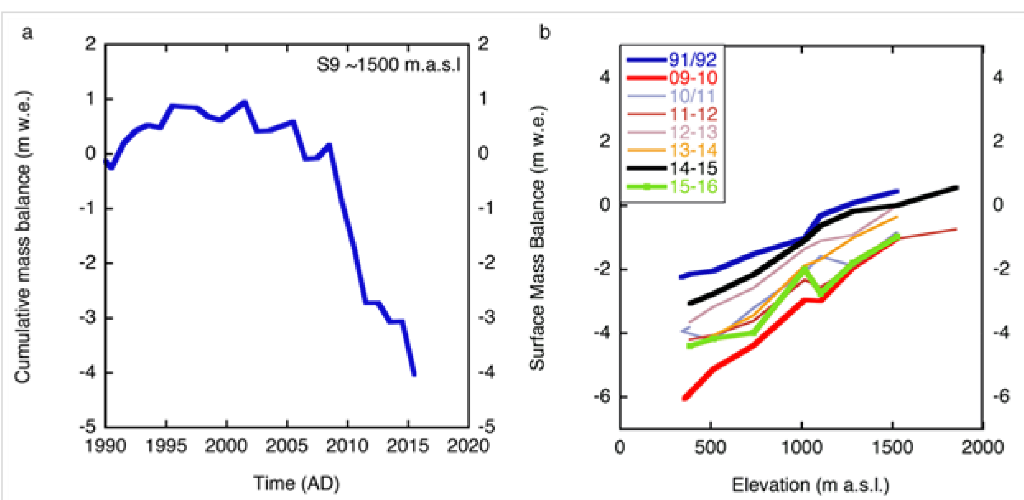

Fig. 3.2. (a) Cumulative mass balance since the start of the measurements in 1991 at S9 (1500 m. a.s.l. and formerly the equilibrium line altitude) along the K-transect. (b) The surface mass balance as a function of elevation along the K-transect for the last seven years. The equilibrium line altitude is defined as the altitude at which SMB = 0 and can be identified as the elevation at which the surface mass balance crosses the zero value.

Net ablation recorded by PROMICE automated stations (www.promice.dk; Van As et al. 2016) for 2016 were all within 1 standard deviation of the 2008-2016 average. Above average melting occurred along the entire ice sheet margin except in the Upernavik (UPE) and Thule (THU) regions, which lie in the northwest (**Fig. 3.3a**). However, when put in a longer-term contest, the same stations show ablation anomalies up to 124% larger than the 1961-1990 average (**Fig. 3.3b**). The largest ablation during the summer of 2016 for the PROMICE stations was 6.3 m w. eq. at the ice sheet margin in the Qassimiut (QAS) region, which is located at the southern portion of the ice sheet, closely followed by the stations of Tasiilaq (TAS) in the southeast and by Kangerlussuaq (KAN) on the west coast.

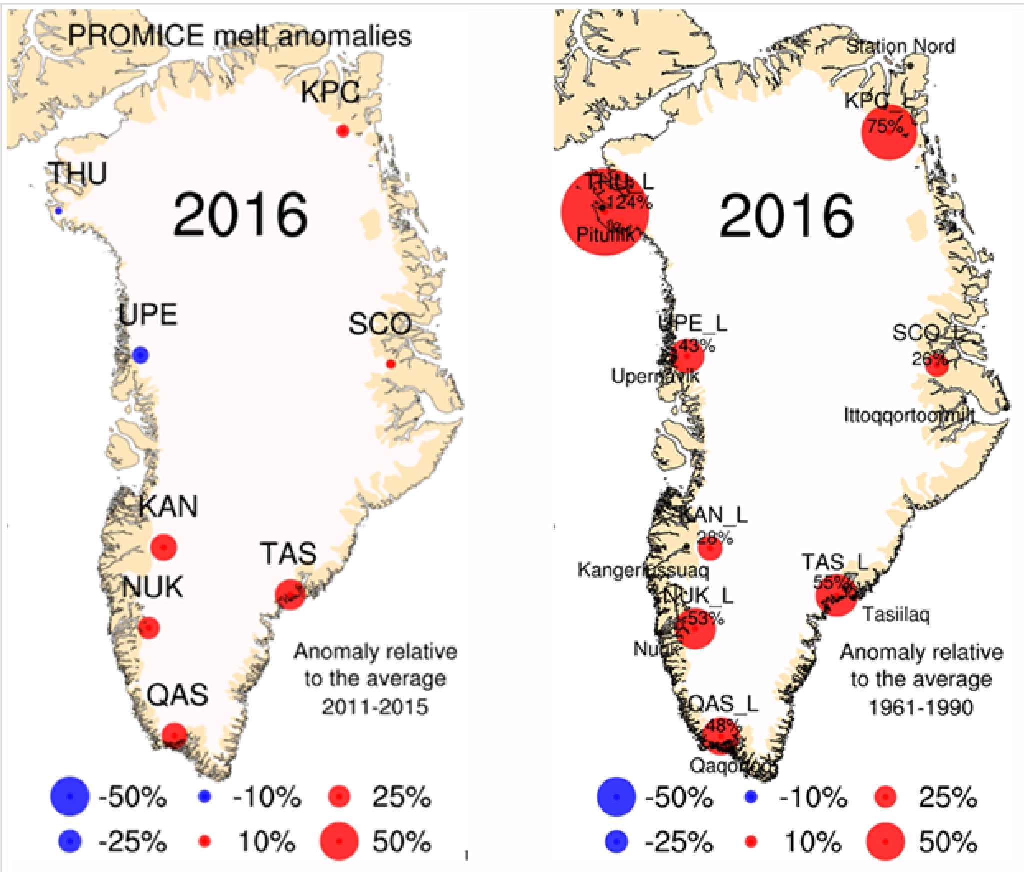

Fig. 3.3. Net ablation anomalies for 2016 at the Programme for Monitoring of the Greenland Ice Sheet (PROMICE) weather station sites, relative to the average over the baseline period (left) 2011-2015 and (right) 1961-1990.

Total Ice Mass

GRACE satellite gravity estimates obtained following Velicogna and Wahr (2014) and available since 2002, indicate that between April 2015 and April 2016 (the most recent period of available data) there was a net ice mass loss of 191 Gt (**Fig. 3.4**). This is about the same as the April 2014-April 2015 mass loss (190 Gt) and smaller than the average April-to-April mass loss (232 Gt) over the period of record. The trend of total ice mass loss for the 14-year is 269 Gt/yr.

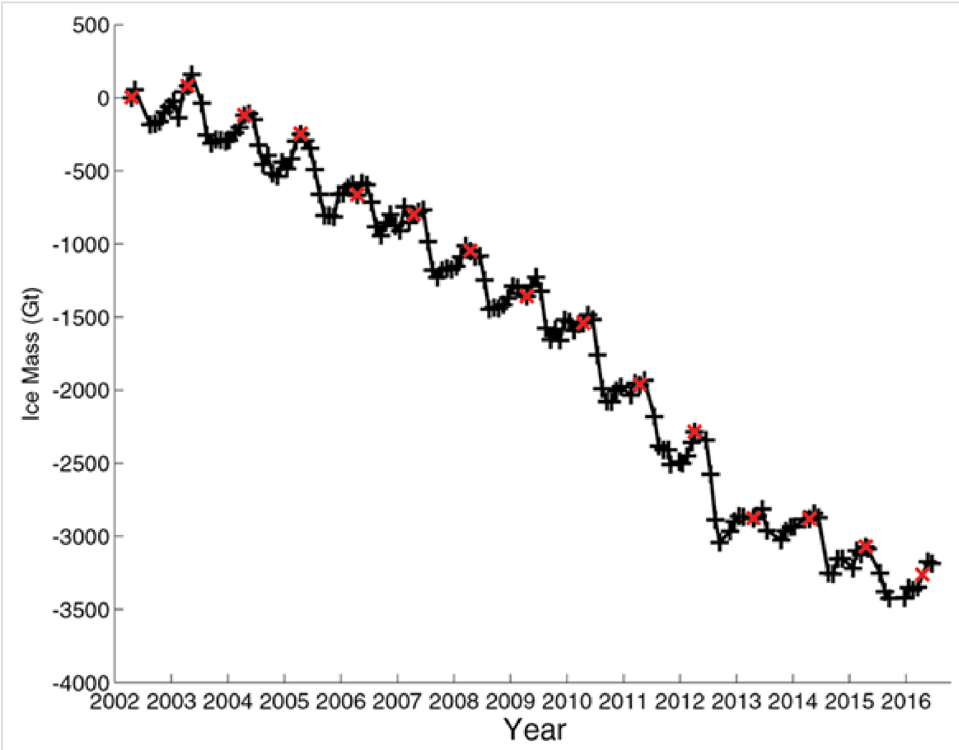

Fig. 3.4. Monthly change in the total mass (in Gigatonnes) of the Greenland ice sheet between April 2002 and April 2016, estimated from GRACE measurements. The red crosses denote the values for the month of April of each year.

Albedo

The average summer albedo is derived from data collected by the Moderate-resolution Imaging Spectroradiometer (MODIS, after Box et al. 2012) and spatially averaged over the entire ice sheet. In 2016, the average summer albedo (June through August, JJA), was 71.1% (**Fig. 3.5a**). This is ~4% lower than in 2000-2001, when MODIS data were first available, and the 5th lowest albedo in the 17 summers of record. The minimum average summer albedo was recorded in 2012 (68.2%), the year of record maximum melt extent. The trends of summer (**Fig. 3.5a**) and July mean (**Fig. 3.5b**) albedo for the period 2000-2016 are, respectively, -5.3±1.0% and -6.2±1.2%. July is the month when the solar incident radiation is at the maximum and, therefore, the impact of albedo in the energy balance is crucial. Low summer albedo anomaly values were widespread (**Fig. 3.5c**). Consistent with the spatial distribution of melt anomalies (**Fig. 3.1b**), the largest area of low albedo anomalies (reaching down to ~ -20 %) was located along the southwestern ice sheet, as in previous recent years (e.g., Tedesco et al. 2016). Only the northwestern portion of the ice sheet was near normal albedo.

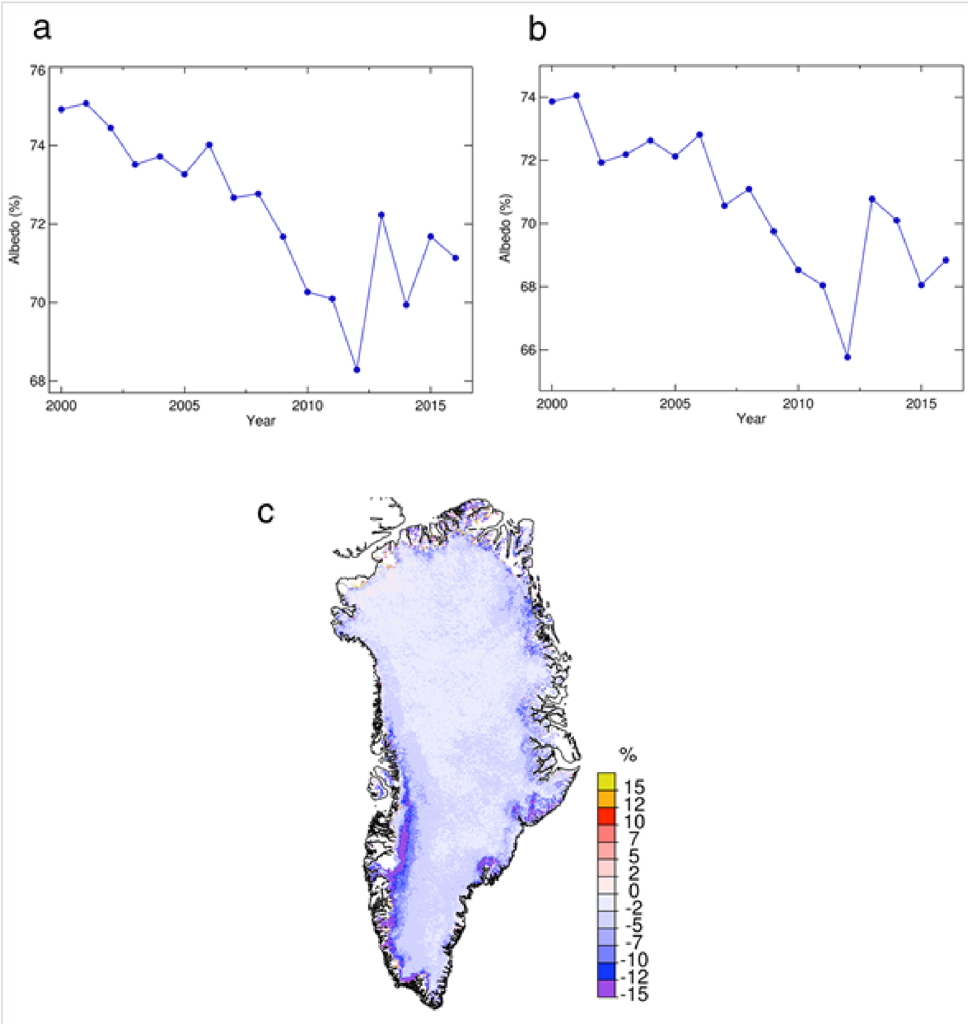

Fig. 3.5. (a) Time series of summer (JJA) and (b) July MODIS albedo averaged over the entire ice sheet. The time series begins in 2000. (c) Map of MODIS albedo anomaly for summer, relative to the 2000-2009 reference period.

Surface Air Temperatures

Also consistent with the spatio-temporal variability of melt and albedo, measurements at fifteen weather stations of the Danish Meteorological Institute (DMI, Cappelen 2016) indicate widespread above average air temperatures in 2016 (relative to the mean for the period 1981-2010). Records were set in 2016 for summer and spring seasons as well as in individual months (**Table 3.1**). New surface temperature records were set during the spring season (March through May, MAM) at Kangerlussuaq, Aasiaat and Summit. April was particularly warm, with new records set at Summit and six other sites. Summer (JJA) temperature anomalies were positive at all stations around the Greenland coastline, with new records set at the southeast coastal site of Tasiilaq, at the northeast site of Danmarkshavn and at the southern site Narsarsuaq. July temperature at Tasiilaq in 2016 was +2.5 °C above normal, second only to 1929, and at Nuuk was second only to 2012. The month of August was record warm at Illoqqortoormiit. At the remaining DMI locations, summer temperatures below the average (e.g., cooler than average).

Table 3.1. Surface temperature anomalies [°C] and z-scores at the different DMI stations for the periods of fall 2015 (SON), winter (DJF) 2015, spring (MAM) 2016 and summer (JJA) 2016. Station names, together with the year in which observations began and the corresponding coordinates are also reported, along with the years when maximum and minimum records were set. Highlighted text indicates the stations and periods when 2016 set a new record.

Station Name, Start Year, Latitude, Longitude		SON 2015	DJF 15/16	MAM 2016	JJA 2016	Station Name, Start Year, Latitude, Longitude		SON 2015	DJF 15/16	MAM 2016	JJA 2016
Pituffik/Thule AFB 1948,76.5,68.8	Anomaly [°C]	0.6	0.3	3.6	1.0	**Narsarsuaq** 1961,61.2, 45.4	Anomaly [°C]	-1.8	0.5	2.8	1.7
	z-score	0	1.9	1	0.5		z-score	0	1.1	**2**	-1.2
	Max Year	2010	1986	1953	1957		Max Year	2010	2010	2010	**2016**
	Min Year	1964	1949	1992	1996		Min Year	1963	1984	1989	1983
Nord 1953, 81.6, 16.7	Anomaly [°C]	0.1	2.5	0.1	0.8	**Quaqortoq** 1873, 60.7,46	Anomaly [°C]	-1.3	1.0	1.8	1.2
	z-score	1.4	0.3	1	0.4		z-score	0.6	0.9	1.3	-0.6
	Max Year	2003	2011	2006	2003		Max Year	2010	2010	1932	1929
	Min Year	1970	1967	1956	1970		Min Year	1874	1884	1989	1874
Upernavik 1873, 72.8, 56.2	Anomaly [°C]	-0.2	0.2	5.4	0.9	**Danmarkshavn** 1949,76.8, 18.8	Anomaly [°C]	2.3	1.8	1.1	2.3
	z-score	0.2	**2.4**	1.3	0		z-score	1	1	**3**	1.7
	Max Year	2010	1947	1932	2012		Max Year	2002	2005	1976	**2016**
	Min Year	1917	1983	1896	1922		Min Year	1971	1967	1966	1955
Kangerlussuaq 1949, 67, 50.7	Anomaly [°C]	-2.7	0.9	6.7	1.3	**Illoqqortoormiut** 1948,70.4,22	Anomaly [°C]	1.1	2.0	2.6	2.2
	z-score	0.1	**2.4**	1.3	-1.6		z-score	1.1	1.7	1.9	1
	Max Year	2010	1986	**2016**	2014		Max Year	2002	2014	1996	1949
	Min Year	1982	1983	1993	1983		Min Year	1951	1966	1956	1955
Ilulissat 1873, 69.2, 51.1	Anomaly [°C]	-2.0	2.5	5.7	0.5	**Tasiilaq** 1895, 65.6, 37.6	Anomaly [°C]	0.6	2.7	2.9	2.3
	z-score	0.8	**2.1**	1.1	-1.3		z-score	1.6	1.8	2.8	0.7
	Max Year	2010	1929	1932	1960		Max Year	1941	1929	1929	**2016**
	Min Year	1884	1884	1887	1972		Min Year	1917	1918	1899	1983
Aasiaat 1951, 68.7, 52.8	Anomaly [°C]	-0.9	3.3	6.0	1.3	**Prins Christian Sund** 1951, 60,43.2	Anomaly [°C]	-0.1	0.4	1.2	0.8
	z-score	0.7	**2.4**	1.2	-0.8		z-score	0.3	1.5	1.1	-0.1
	Max Year	2010	2010	**2016**	2012		Max Year	2010	2010	2005	2010
	Min Year	1986	1984	1993	1972		Min Year	1982	1993	1989	1970
Nuuk 1873, 64.2, 51.8	Anomaly [°C]	-1.6	0.5	3.8	2.4	**Summit** 1988, 72.6, 38.5	Anomaly [°C]	0.4	-0.5	4.8	1.1
	z-score	0.3	**2.1**	**2.3**	-1.3		z-score	-0.2	2.2	0.7	0.2
	Max Year	2010	2010	1932	2012		Max Year	1987	2010	**2016**	2012
	Min Year	1898	1984	1993	1914		Min Year	2009	1993	1992	1992
Paamiut 1958,62,49.7	Anomaly [°C]	-1.0	1.9	2.4	0.7						
	z-score	0.5	1.3	0.8	-0.9						
	Max Year	2010	2010	2005	2010						
	Min Year	1982	1984	1993	1969						
	Min Year	2009	1993	1992	1992						

Data recorded by the PROMICE network (sites shown in **Fig. 3.3**) are consistent, indicating above-average air temperatures for 88% of all January-August 2016 station-months.

The average Greenland Blocking Index (GBI, here defined as the average 500h Pa geopotential height for the region 60°-80° N and 20-80° W, e.g., Hanna et al. 2013) calculated from the NCAR/NCEP Reanalysis was the second highest since 1948, following only the extensive melt year of 2012 (Nghiem et al. 2012). Persistent periods of high GBI values have been associated with extensive Greenland surface melt and negative surface mass balance (Hanna et al. 2013, McLeod and Mote 2015). Despite the near record GBI, the average daily melt during summer of 2016 was much less than the record breaking year of 2012. A major difference between the 2012 and 2016 atmospheric conditions was the lack of water vapor transport, and associated latent heat and downwelling longwave radiative fluxes in 2016, which have recently been shown to have a considerable effect on ice sheet melt (Mattingly et al. 2016). The average daily melt during summer of 2016 was comparable to 2015 and 2014, continuing the overall trend towards increasing melting.

Marine-terminating Glaciers

Marine-terminating glaciers are the outlets via which the Greenland ice sheet discharges ice mass to the ocean. When in balance, the rate of iceberg calving (by area) is equalized by the seaward flow of the ice (see the essay on Greenland Ice Sheet Surface Velocity: New Data Sets, 2015 Arctic Report Card for further information on the velocity of these glaciers). Glacier area measurements from optical satellite imagery (Landsat and ASTER) area available since 1999. These measurements evaluated at the end of the melt season to minimize the influence of observed winter advance (Jensen et al. 2016). In the most recent year, from 2015 to 2016, the area changes at 45 glaciers that are among the widest and fastest-flowing marine-terminating indicate a continued pattern of retreat.

With a total net glacier front area change of -60.6 km^2 (net area loss), 2016 was characterized by the largest area loss of marine terminating glaciers since 2012 (**Fig. 3.6**). A total of 22 glaciers retreated (100.8 km^2 area loss), 11 glaciers advanced (40.9 km^2 area gain), and 12 remained relatively stable, with a net area change smaller than 0.2 km^2. The area changes do not count the -100 km^2 area from Spalte glacier that detached after years of rift propagation. The largest area losses occurred at northern Greenland glaciers, led by Ryder (-23.5 km^2), Zachariae (-15.5 km^2) and Storstrommen (-11.0 km^2) glaciers. The largest advances were observed at Jakobshavn (+4.5 km^2), Helheim (+2.1 km^2), and Upernavik A (+1.6 km^2). See http://gac.geus.dk/ for glacier locations.

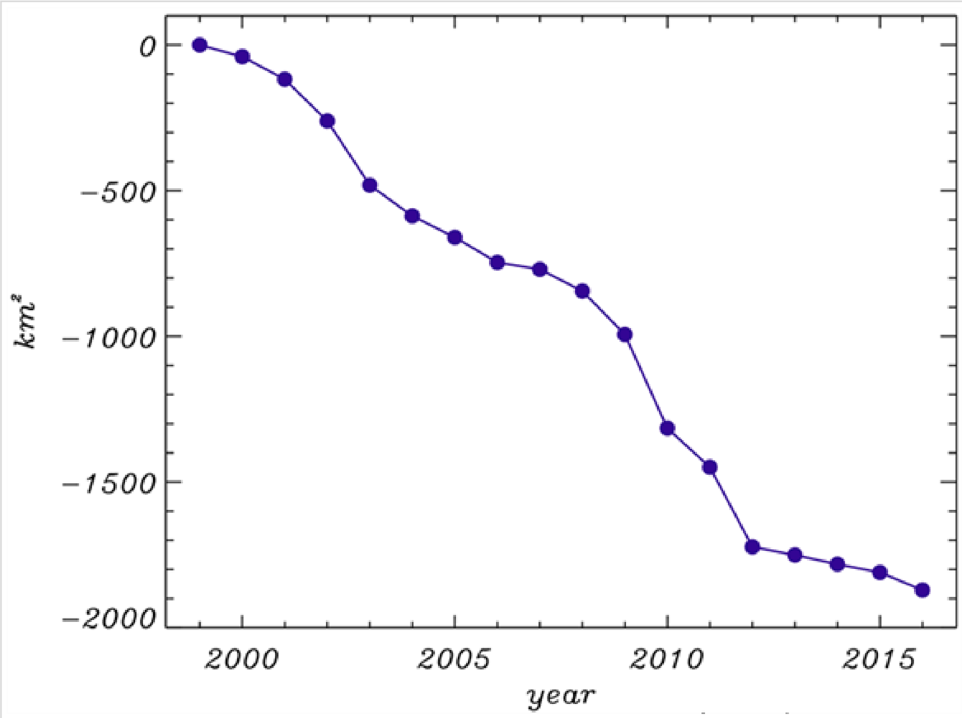

Fig. 3.6. Cumulative net area change (km^2 and miles^2, left and right axes, respectively) at the 45 of the widest and fastest-flowing marine-terminating glaciers of the Greenland Ice Sheet (after Jensen et al. 2015 and Box and Hansen 2016).

Acknowledgments

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References

Box, J. E. and K. Hansen, 2015. Survey of Greenland glacier area changes, PROMICE newsletter 8, December 2015, http://promice.org/Newsletter_08.pdf.

Box, J. E., X. Fettweis, J. C. Stroeve, M. Tedesco, D. K. Hall, and K. Steffen, 2012. Greenland ice sheet albedo feedback: thermodynamics and atmospheric drivers. *The Cryosphere*, 6, 821-839, doi:10.5194/tc-6-821-2012.

Cappelen (ed), 2016. Greenland - DMI Historical Climate Data Collection 1784-2015. *Danish Meteorol. Inst. Rep.*, 16-04.

Hanna, E., J. M. Jones, J. Cappelen, S. H. Mernild, L. Wood, K. Steffen, and P. Huybrechts, 2013. The influence of North Atlantic atmospheric and oceanic forcing effects on 1900-2010 Greenland summer climate and ice melt/runoff. *Int. J. Climatol*, 33, 862-880, doi: 10.1002/joc.3475.

Jensen, T., J. E. Box, and C. Hvidberg, 2016. A sensitivity study of yearly Greenland ice sheet marine terminating outlet glacier changes: 1999-2013, *Journal of Glaciology*, 62(231), 72-81.

Mattingly, K., C. Ramseyer, J. Rosen, T. Mote, and R. Muthyala, 2016. Increasing water vapor transport to the Greenland Ice Sheet revealed using self-organizing maps, *Geophys. Res. Lett.*, doi: 10.1002/2016GL070424.

McLeod, J. T., and T. L. Mote, 2015. Linking interannual variability in extreme Greenland blocking episodes to the recent increase in summer melting across the Greenland ice sheet. *Int. J. Climatol.*, 36, 1484-1499, doi: 10.1002/joc.4440.

Mote, T., 2007. Greenland surface melt trends 1973-2007: Evidence of a large increase in 2007. *Geophys. Res. Lett.*, 34, L22507.

Nghiem, S. V., D. K. Hall, T. L. Mote, M. Tedesco, M. R. Albert, K. Keegan, C. A. Shuman, N. E. DiGirolamo, and G. Neumann, 2012. The extreme melt across the Greenland ice sheet in 2012, *Geophys. Res. Lett.*, 39, L20502, doi:10.1029/2012GL053611.

Oerlemans, J., 2012. Twenty-one years of mass balance observations along the K-transect, West-Greenland. *Earth Syst. Sci. Data*, 4, 31-35, doi:10.5194/essd-4-31-2012.

Tedesco, M., X. Fettweis, T. Mote, J. Wahr, P. Alexander, J. Box, and B. Wouters, 2013. Evidence and analysis of 2012 Greenland records from spaceborne observations, a regional climate model and reanalysis data. *The Cryosphere*, 7, 615-630.

Tedesco, M. et al., 2016. Greenland Ice Sheet [in "State of the Climate in 2015"]. *Bull. Amer. Meteor. Soc.*, 97 (8), S140-S142.

Van As, D., Fausto, R. S., Cappelen, J., Van de Wal, R. S. W., Braithwaite, R. J., Machguth, H., and PROMICE project team, 2016. Placing Greenland ice sheet ablation measurements in a multi-decadal context. *Geol. Surv. Denmark Greenland Bull.*, 35, 71-74.

Van de Wal, R. S. W., W. Greuell, M. R. van den Broeke, C. H. Reijmer, and J. Oerlemans, 2005. Surface mass-balance observations and automatic weather station data along a transect near Kangerlussuaq, West Greenland. *Ann. Glaciol.*, 42, 311-316.

Van de Wal, R. S. W., W. Boot, C. J. P. P. Smeets, H. Snellen, M. R. van den Broeke, and Velicogna, I. and J. Wahr, 2006. Significant acceleration of Greenland ice mass loss in spring, 2004. *Nature*, 022 Sep 2006; doi:10.1038/nature05168.

November 15, 2016

Sea Ice

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Highlights

- The September 2016 Arctic sea ice minimum extent was 4.14 million km^2, 33% lower than the 1981-2010 average minimum ice extent and tied with 2007 for the second lowest value in the satellite record (1979-2016).
- The lowest winter maximum ice extent in the satellite record (1979-2016) occurred on 24 March 2016, at 14.52 million km^2, 7% below the 1981-2010 average.
- In March 2016, multiyear ice (more than 1 year old) and first-year ice were 22% and 78% of the ice cover, respectively, compared to 45% and 55% in 1985.

Sea Ice Extent

The Arctic sea ice cover is vast in areal extent covering millions of square kilometers, but is only a thin veneer a few meters thick. This ice cover plays many roles. It is a barrier limiting the exchange of heat, moisture, and momentum between the atmosphere and ocean; a home to a rich marine ecosystem, including human communities; and an indicator of climate change. Sea ice extent has been monitored using passive microwave instruments on satellite platforms since 1979. The months of September and March are of particular interest because they are the months when the Arctic sea ice typically reaches its maximum and minimum extent respectively. Maps of monthly average ice extents in March 2016 and September 2016 are shown in **Fig. 4.1**. The major difference in March 2016 compared to the 1981-2010 average was a large ice-free area north of Svalbard and Novaya Zemlya.

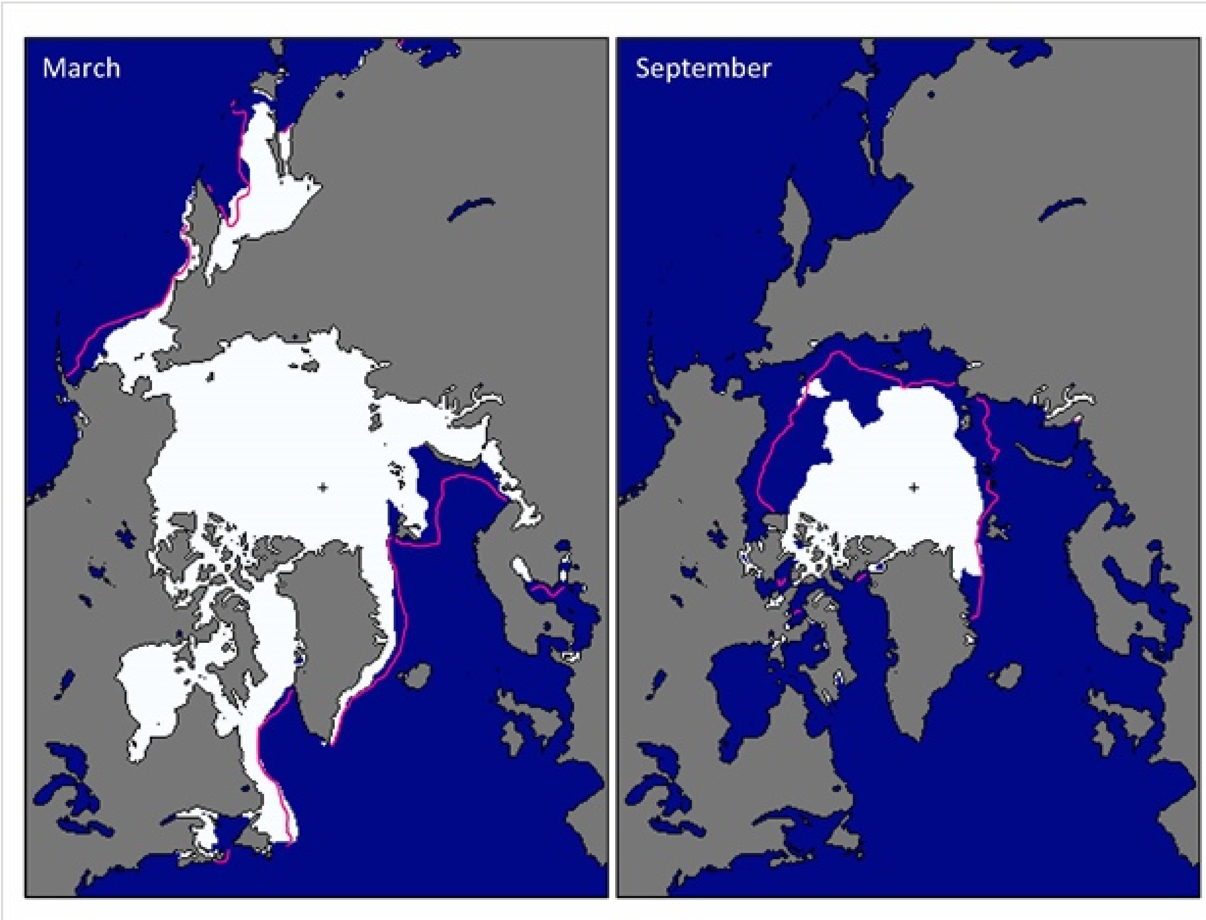

Fig. 4.1. Average monthly sea ice extent in March 2016 (left) and September 2016 (right) illustrate the respective winter maximum and summer minimum extents. The colored line indicates the median ice extents in March and September, respectively, during the period 1981-2010. Maps are from NSIDC at nsidc.org/data/seaice_index (Fetterer et al., 2002).

Based on estimates produced by the National Snow and Ice Data Center (NSIDC) Sea Ice Index (Fetterer et al., 2002), the sea ice cover reached a minimum annual extent of 4.14 million km^2 on September 10, 2016. This matched 2007 as the second lowest minimum extent in the satellite record. The 2016 summer minimum extent is larger by 0.75 million km^2 (22%) than the record minimum of 3.39 million km^2 set in 2012. It was, however, 1.81 million km^2 (29%) less than the 1981-2010 average minimum ice extent and 0.29 million km^2 (6.5%) less than the 2015 minimum.

On March 24, 2016 ice extent reached a winter maximum value of 14.52 million km^2, 7% below the 1981-2010 average. This matched 2015 as the lowest maximum value in the satellite record. Also notable, the maximum extent occurred 12 days later than the 1981-2010 average (12 March) and was the fourth latest of the satellite record. The date of the maximum has been trending slightly later (1.7 days per decade) over the satellite record, albeit with considerable year-to-year variability.

Sea ice extent has decreasing trends in all months and virtually all regions, the exception being the Bering Sea during winter (Meier et al., 2014). The September monthly average trend for the entire Arctic Ocean is now -13.3% per decade relative to the 1981-2010 average (**Fig. 4.2**). While the 2016 daily minimum ice extent was the second lowest on record, the monthly value (shown in **Figure 4.2**) was only the fifth lowest due to rapid ice formation in late September. Trends are smaller during March (-2.7% per decade), but are still decreasing at a statistically significant rate.

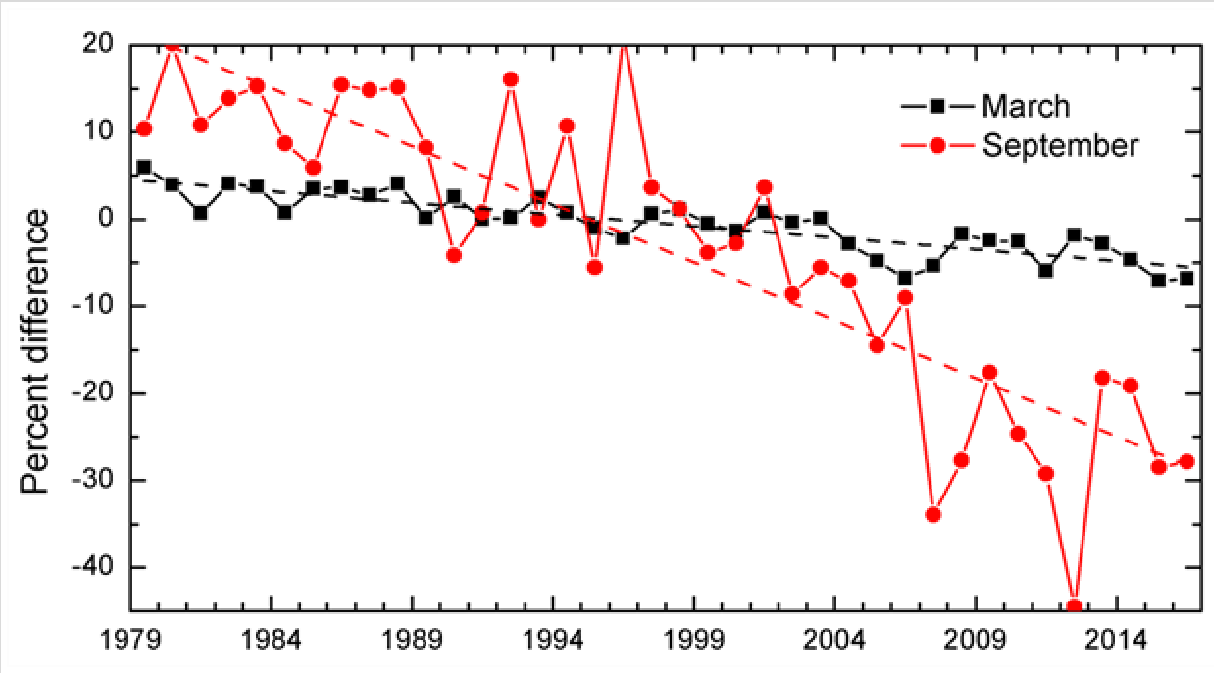

Fig. 4.2. Time series of ice extent anomalies in March (the month of maximum ice extent) and September (the month of minimum ice extent). The anomaly value for each year is the difference (in %) in ice extent relative to the mean values for the period 1981-2010. The black and red lines are least squares linear regression lines. The slopes of these lines indicate ice loss of 2.7% and 13.3% per decade in March and September, respectively. Both trends are significant at the 99% confidence level.

In 2016 year, 10.38 million km^2 of ice was lost between the March maximum and September minimum extent. Before 2007, a March to September loss of more than 10 million km^2 of ice occurred only once (1991), but since 2007 such large losses have occurred 7 out of 10 years.

Age of the Ice

The age of sea ice is another key descriptor of the state of the sea ice cover. It is an indicator for ice physical properties, including surface roughness, melt pond coverage and thickness. Older ice tends to be thicker and thus more resilient to changes in atmospheric and oceanic forcing than younger ice. The age of the ice is determined using satellite observations and drifting buoy records to track ice parcels over several years (Tschudi et al. 2010; Maslanik et al. 2011). This method has been used to provide a record of the age of the ice since the early 1980s (Tschudi et al. 2015).

The oldest ice (>4 years old) continues to make up a small fraction of the Arctic ice pack in March, when sea ice is at its maximum annual extent (**Fig. 4.3**). In 1985, 16% of the ice pack (relative to the total sea ice areal coverage) was four years old and older, but by March 2016 old ice only constituted 1.2% of the ice pack. First-year ice now dominates the ice cover, comprising about 78% of the March 2016 ice pack, compared to about 55% in the 1980s. The distribution of ice age in March 2016 was similar to that in March 2015. Given that older ice tends to be thicker, the sea ice cover has transformed from a strong, thick pack in the 1980s to a more fragile, younger, and thinner pack in recent years. The thinner, younger ice is more vulnerable to melting out in the summer, contributing to lower minimum ice extents.

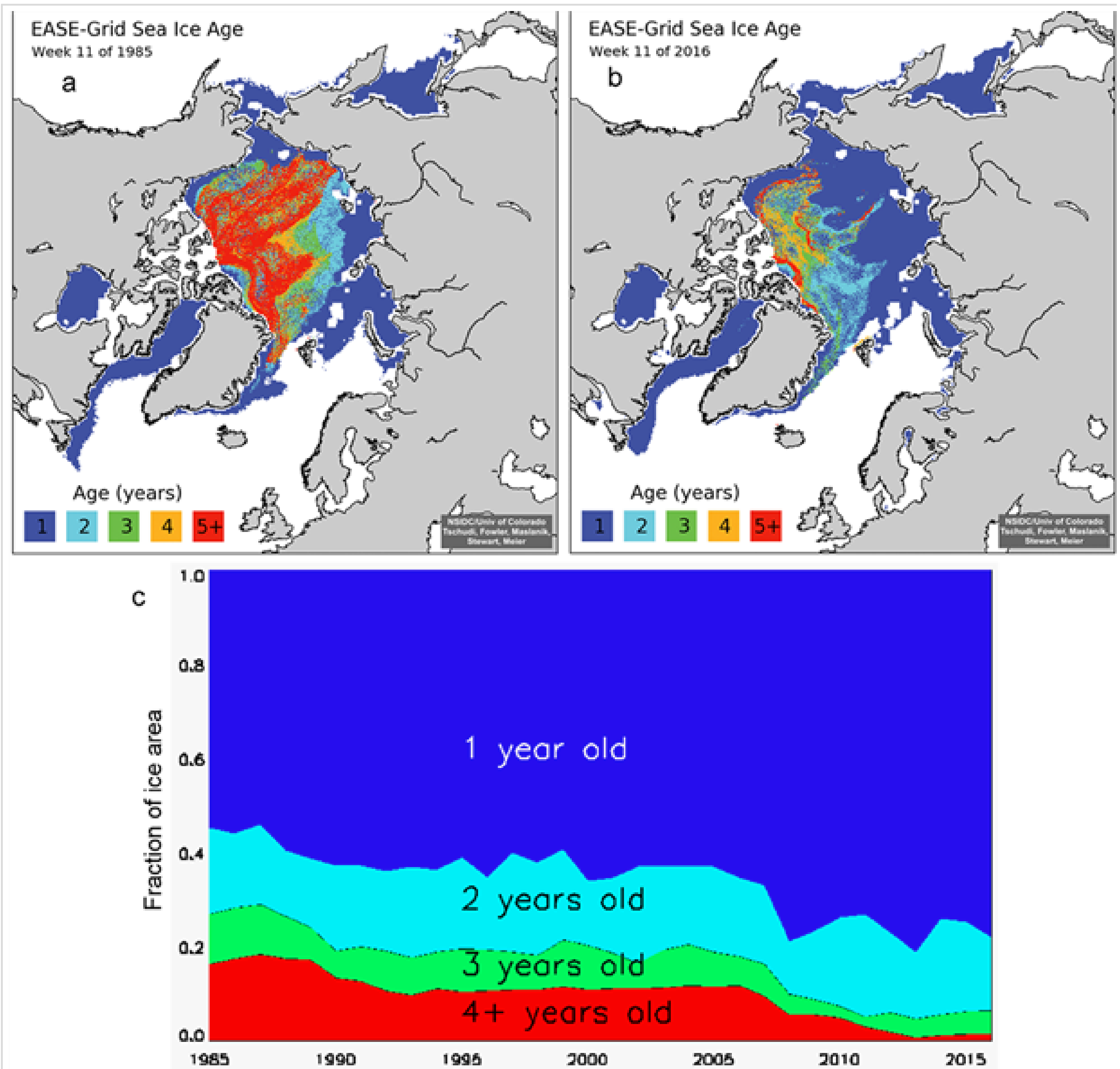

Fig. 4.3. Sea ice age coverage maps for a) March 1985 (Tschudi et al., 2015), b) March 2016 (personal communication, J. S. Stewart), and c) time series of sea ice age coverage, 1985-2016 (provided by M. Tschudi). The coverages in (c) are presented as fractions, or percent, of the total sea ice areal coverage.

Sea Ice Thickness

Observations of sea ice thickness and volume from multiple sources have revealed the continued decline of the Arctic sea ice pack over the last decade (Kwok and Rothrock 2009; Laxon et al. 2013; Kwok and Cunningham 2015; Lindsay and Schweiger, 2015). These changes have impacts on the regional Arctic and sub-Arctic climate, environment and ecosystems. To understand these impacts, as the Arctic transitions from a predominantly multi-year ice pack to a seasonal ice cover (**Fig. 4.3**), continued monitoring of the thickness of the ice pack is required. The European Space Agency CryoSat-2 satellite has been measuring sea ice freeboard (from which sea ice thickness and volume are derived) since 2010 (Tilling et al., 2015). Measurements of sea ice plus snow thickness by airborne electromagnetic induction sounding (AEM) from helicopters and aircrafts by various agencies have been made during late summer since 2001 (Haas et al. 2010).

CryoSat-2 derived Arctic sea-ice thickness in April 2016 are presented in **Fig. 4.4**. The ice is near its maximum annual thickness in April at the end of winter. Also plotted is the 2016 anomaly compared to the average April values from 2011 to 2015. As in previous years, results show a thickness gradient across the central Arctic Ocean between the oldest, thickest ice near Greenland and the Canadian Arctic Archipelago (3-4 m), and younger ice in the Beaufort, East Siberian and Laptev Seas (≤2 m) (**Fig. 4.4a**). In the context of the average conditions, the April 2016 results (**Fig. 4.4b**) show two distinct features. The first was a band of thick multi-year ice in the southern and eastern Beaufort Sea that is surrounded by thinner ice (up to 1 m below average) in the western Beaufort Sea and Canada Basin. The second feature was a region with above average thickness north of Fram Strait in the Eurasian Basin. Airborne electromagnetic induction surveys (York University) obtained independent measurements of ice thickness in this region and confirmed the CryoSat-2 results. Such spatial and temporal variability in ice thickness results from variability in the motion of the ice. Areas of ice divergence typically have thinner ice while convergence ice causes ridging and thicker ice.

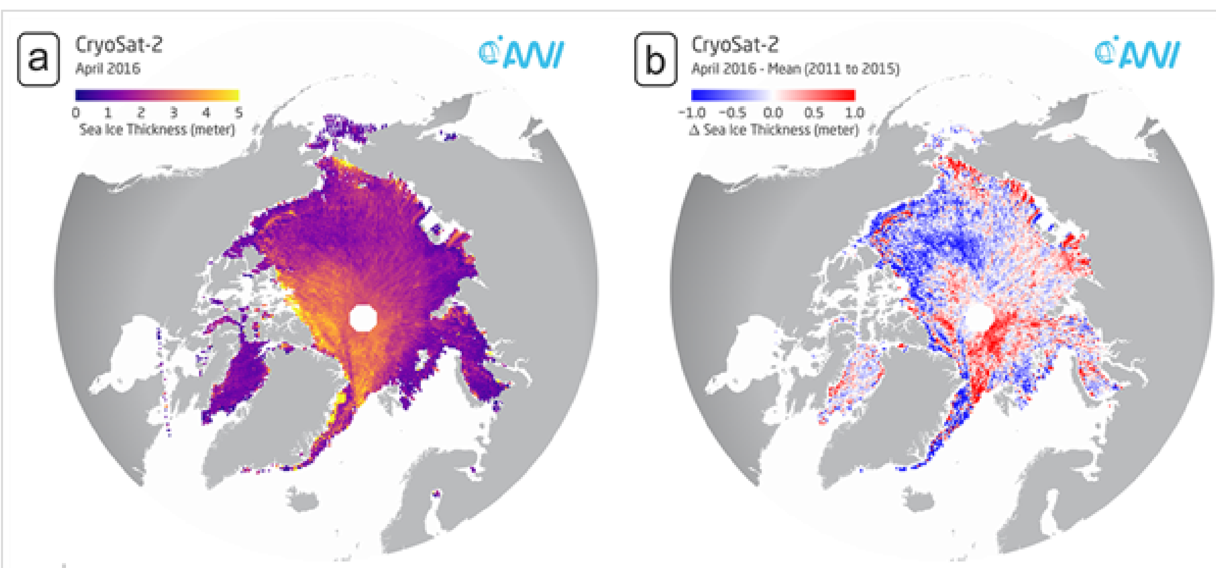

Fig. 4.4. Observations of sea ice thickness. (a) Sea ice thickness derived from ESA CryoSat-2 in April 2016. (b) Sea ice thickness anomaly in April 2016 to the mean of all previous years (2011 - 2015) of the CryoSat-2 observational data record. Blue indicates regionally thinner and red indicates thicker sea ice in 2016 than the 5 year average.

References

Fetterer, F., K. Knowles, W. Meier, and M. Savoie. 2002, updated daily. *Sea Ice Index*. Boulder, Colorado USA: National Snow and Ice Data Center. http://dx.doi.org/10.7265/N5QJ7F7W.

Haas, C., Hendricks S., Eicken H., Herber A.: Synoptic airborne thickness surveys reveal state of Arctic sea ice cover. *Geophysical Research Letters* 05/2010; 37(9). doi:10.1029/2010GL042652.

Kwok, R., and D. A. Rothrock, 2009: Decline in Arctic sea ice thickness from submarine and ICESat records: 1958-2008, *Geophys. Res. Lett.*, 36, doi:10.1029/2009GL039035.

Kwok, R., and G. F. Cunningham, 2015: Variability of Arctic sea ice thickness and volume from CryoSat-2, *Phil. Trans. Royal Soc. London A: Math., Phys. Eng. Sci.*, 373, 2045, doi.org/10.1098/rsta.2014.0157.

Laxon, S. W., K. A. Giles, A. L. Ridout, D. J. Wingham, R. Willatt, R. Cullen, R. Kwok, A. Schweiger, J. Zhang, C. Haas, S. Hendricks, R. Krishfield, N. Kurtz, S. L. Farrell, M. Davidson, 2013: CryoSat estimates of Arctic sea ice Volume, *Geophys. Res. Lett.*, 40, doi:10.1002/grl.50193.

Lindsay, R. and A. Schweiger, 2015. Arctic sea ice thickness loss determined using subsurface, aircraft, and satellite observations. *The Cryosphere*, 9:269-283.

Maslanik, J., J. Stroeve, C. Fowler, and W. Emery, 2011: Distribution and trends in Arctic sea ice age through spring 2011. *Geophys. Res. Lett.*, 38, doi: 10.1029/2011GL047735.

Meier, W. N., G. Hovelsrud, B. van Oort, J. Key, K. Kovacs, C. Michel, M. Granskog, S. Gerland, D. Perovich, A. P. Makshtas, and J. Reist, 2014. Arctic sea ice in transformation: A review of recent observed changes and impacts on biology and human activity, *Rev. Geophys.*, 41, doi:10.1002/2013RG000431.

NSIDC at nsidc.org/data/seaice_index. Colorado USA. NASA National Snow and Ice Data Center Distributed Active Archive Center, doi.org/10.5067/1UQJWCYPVX61.

Tilling, R. L., A. Ridout, A. Shepherd, and D. J. Wingham, 2015; Increased Arctic sea ice volume after anomalously low melting in 2013. *Nat. Geosci.*, 8, 643-646. doi:10.1038/ngeo2489.

Tschudi, M. A., C. Fowler, J. A. Maslanik, and J. A. Stroeve, 2010: Tracking the movement and changing surface characteristics of Arctic sea ice. *IEEE J. Sel. Topics Earth Obs. and Rem. Sens.*, 3, doi: 10.1109/JSTARS.2010.2048305.

Tschudi, M., C. Fowler, and J. Maslanik, 2015: EASE-Grid Sea Ice Age, Version 2. Boulder, Colorado USA. NASA National Snow and Ice Data Center Distributed Active Archive Center, doi.org/10.5067/1UQJWCYPVX61.

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Sea Surface Temperature

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Highlights

- Sea surface temperatures (SSTs) in August 2016 were up to +5° C warmer than the 1982-2010 August mean in regions of the Barents and Chukchi seas, and off the east and west coasts of Greenland.
- In the Arctic Basin, spatial patterns of August 2016 SST anomalies relative to the 1982-2010 August mean are linked to regional variability in sea-ice retreat, regional air temperature, and advection of waters from the Pacific and Atlantic oceans.
- The Chukchi Sea and eastern Baffin Bay show significant ocean surface warming trends; linear trends over 1982-2016 indicate August SSTs are increasing at ~0.5° C/decade in these regions.

Summer sea surface temperatures in the Arctic Ocean are set by absorption of solar radiation into the surface layer. In the Barents and Chukchi seas, there is an additional contribution from advection of warm water from the North Atlantic and Pacific Oceans (for a recent assessment of this in the Chukchi Sea, see Serreze et al., 2016). Solar warming of the ocean surface layer is influenced by the distribution of sea ice (with more solar warming in ice-free regions), cloud cover, water color, and upper-ocean stratification (river influxes influence the latter two). Here August SSTs are reported, which are an appropriate representation of Arctic Ocean summer SSTs and are not affected by the cooling and subsequent sea-ice growth that takes place in the latter half of September. SST data are from the NOAA Optimum Interpolation (OI) SST Version 2 product (OISSTv2), which is a blend of in situ and satellite measurements (Reynolds et al. 2002, 2007). Compared to in situ temperature measurements, the OISSTv2 product showed average correlations of about 80%, with an overall cold SST bias of -0.02° C (Stroh et al., 2015).

Mean SSTs in August 2016 in ice-free regions ranged from ~0° C in some regions to around +7° C to +8° C in the Chukchi Sea and eastern Baffin Bay off the west coast of Greenland, and up to +11° C in the Barents Sea (**Fig. 5.1a**). Most boundary regions and marginal seas of the Arctic had anomalously warm SSTs in August 2016 compared to the 1982-2010 August mean (**Fig. 5.1b**). SSTs in these regions, which are mostly ice-free in August, are linked to the timing of local sea-ice retreat; for example, anomalously warm SSTs (>+2° C relative to 1982-2010) in August 2016 in the Beaufort Sea may be associated with low sea-ice extents and exposure of surface waters to direct solar heating. Compared to August 2012 (the summer of lowest minimum sea-ice extent in the satellite record, 1979-present) Beaufort Sea SSTs were up to 2° C cooler in 2016 (**Fig. 5.1c**). August 2016 SSTs were cooler relative to the 1982-2010 average along the southern boundaries of the Beaufort Sea and East Siberian and Laptev Seas (**Fig. 5.1b**), where summer air temperatures were also below average (see essay on *Surface Air Temperature*); cooler than average SSTs were also notable in the northern Barents Sea. August 2016 SST anomalies off the east and west coasts of Greenland, and in the southern Barents Sea were up to 5° C warmer than the 1982-2010 average, and coincide with regional surface air temperatures that were up to 5° C warmer in July-August 2016 compared to 1982-2010 July-August mean temperatures (see essay on

Surface Air Temperature); SSTs in these regions were up to 3° C warmer in August 2016 compared to August 2012 (**Fig. 5.1c**).

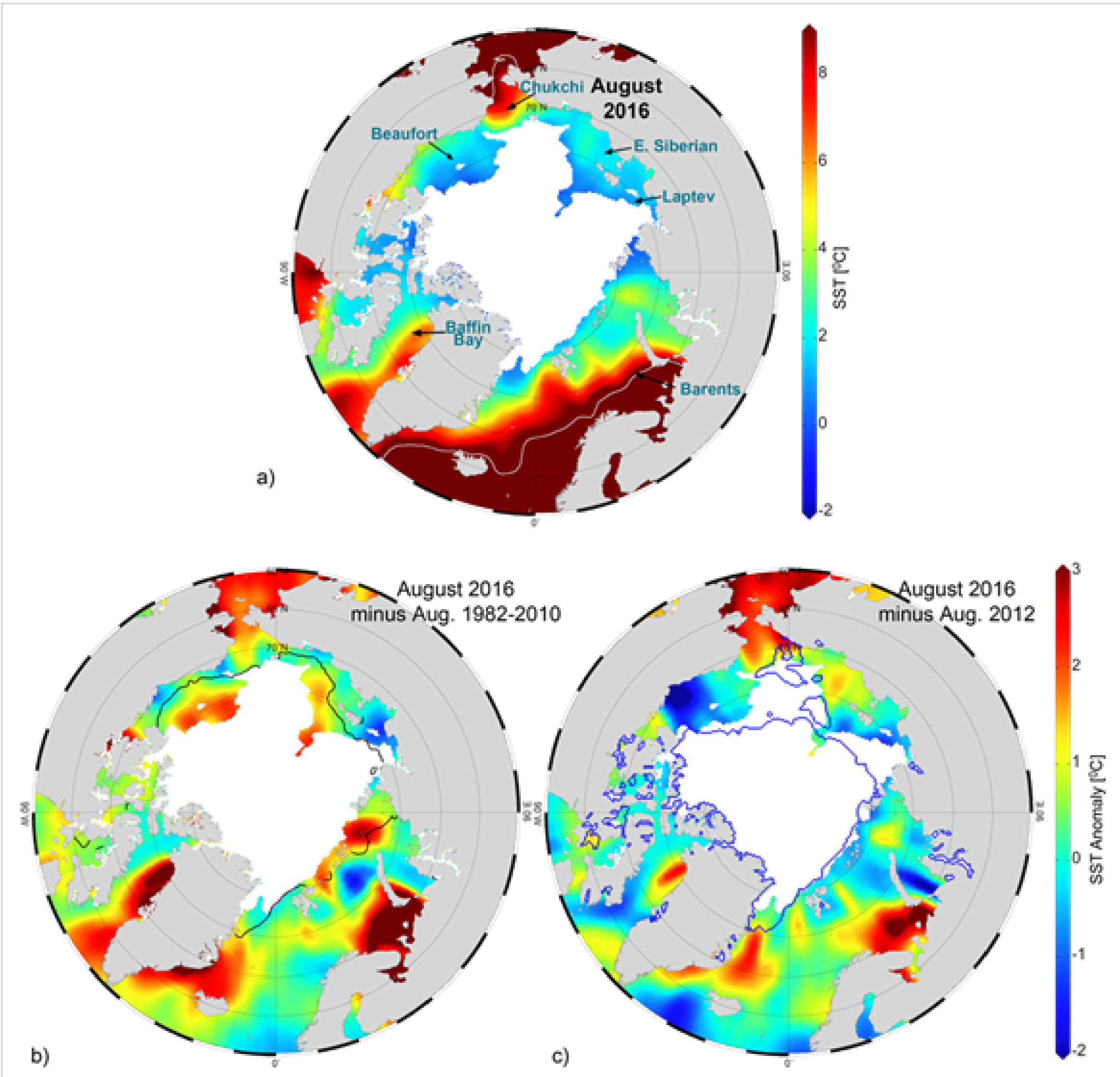

Fig. 5.1. (a) Mean sea surface temperature (SST, °C) in August 2016. White shading is the August 2016 mean sea ice extent, and gray contours indicate the 10° C SST isotherm. (b) SST anomalies (°C) in August 2016 relative to the August mean for the period 1982-2010. White shading is the August 2016 mean ice extent and the black line indicates the median ice edge in August for the period 1982-2010. (c) SST anomalies (°C) in August 2016 relative to August 2012 (the year of lowest minimum sea-ice extent in the satellite record: 1979-present); white shading is the August 2016 mean ice extent and the blue line indicates the median ice edge for August 2012. Sea-ice extent and ice-edge data are from NSIDC.

The Chukchi Sea and eastern Baffin Bay are the only marginal regions to exhibit a statistically significant warming trend over the duration of the record (August SSTs in these regions are warming at a rate of about

+0.5° C decade^{-1} since 1982, based on a linear fit; **Fig. 5.2**). In the Chukchi Sea, this trend coincides with declining trends in summer sea ice extent. In other marginal seas, for instance the Barents Sea, warm August SST anomalies observed in 2016 are of similar magnitude to warm anomalies observed in past decades (Timmermans and Proshutinsky 2015; 2016).

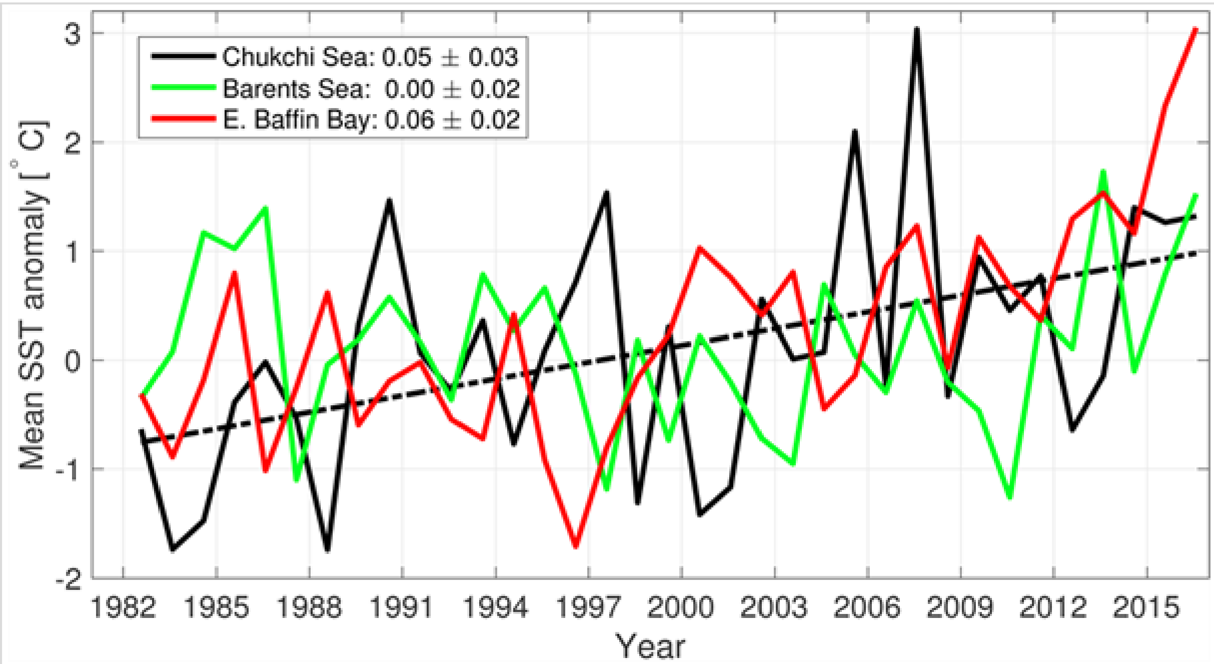

Fig. 5.2. Time series of area-averaged SST anomalies (°C) for August of each year relative to the August mean for the period 1982-2010 for the Chukchi and Barents seas and eastern Baffin Bay (see **Fig. 1a**). The dash-dotted black line shows the linear SST trend (over the period shown) for the Chukchi Sea. Numbers in the legend correspond to linear trends in °C year^{-1} (with 95% confidence intervals).

References

Reynolds, R. W., N. A. Rayner, T. M. Smith, D. C. Stokes, and W. Wang, 2002: An improved in situ and satellite SST analysis for climate. *J. Climate*, 15, 1609-1625.

Reynolds, R. W., T. M. Smith, C. Liu, D. B. Chelton, K. S. Casey, and M. G. Schlax, 2007: Daily high-resolution-blended analyses for sea surface temperature. *J. Climate*, 20, 5473-5496, and see http://www.esrl.noaa.gov/psd/data/gridded/data.noaa.oisst.v2.html.

Serreze, M. C., A. D. Crawford, J. Stroeve, A. P. Barrett and R. A. Woodgate, 2016: Variability, trends and predictability of seasonal sea ice retreat and advance in the Chukchi Sea. *J. Geophys. Res.*, accepted for publication.

Stroh, J. N., G. Panteleev, S. Kirillov, M. Makhotin, and N. Shakhova, 2015: Sea-surface temperature and salinity product comparison against external in situ data in the Arctic Ocean. *J. Geophys. Res. Oceans*, 120, 7223-7236, doi:10.1002/2015JC011005.

Timmermans, M. -L. and A. Proshutinsky, 2015: [The Arctic] Sea Surface Temperature [in "State of the Climate in 2014"]. *Bull. Amer. Meteor. Soc.*, 96 (7), S147-S148.

Timmermans, M. -L. and A. Proshutinsky, 2016: [The Arctic] Sea Surface Temperature [in "State of the Climate in 2015"]. *Bull. Amer. Meteor. Soc.*, 97 (8), S137-S138.

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Arctic Ocean Primary Productivity

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Highlights

- During May 2016, chlorophyll-*a* concentrations averaged more than ~14 mg m^{-3} higher than the 2003-2015 mean across a relatively large (~300 x 300 km) region in the central Barents Sea. During June 2016 a ~150 x 750 km band in the Greenland Sea northeast of Greenland also experienced relatively high chlorophyll-*a* concentrations near the sea ice edge, where values were ~7 mg m^{-3} higher than the 2003-2015 mean.
- The steepest trends in chlorophyll-*a* concentrations over the years 2003-2016 have occurred during May in localized areas of the Barents Sea, with an overall positive trend averaging ~0.79 mg m^{-3} yr^{-1}.
- Estimates of ocean primary productivity showed widespread positive anomalies for 2016, except for the western (North American) Arctic and the Sea of Okhotsk.

Introduction

Primary productivity is the rate at which atmospheric or aqueous carbon dioxide is converted by autotrophs (primary producers) to organic material. Primary production via photosynthesis is a key process within the ecosystem, as the producers form the base of the entire food web, both on land and in the oceans. The oceans play a significant role in global carbon budgets via photosynthesis. Approximately half of all global net annual photosynthesis occurs in the oceans, with ~10-15% of production occurring on the continental shelves alone (Müller-Karger et al. 2005). Primary productivity is strongly dependent upon light availability and the presence of nutrients, and thus is highly seasonal in the Arctic region. In particular, the melting and retreat of sea ice during spring are strong drivers of primary production in the Arctic Ocean and its adjacent shelf seas due to enhanced light availability (Barber et al. 2015, Leu et al. 2015). Recent declines in Arctic sea ice extent (see the essay on *Sea Ice*) have contributed substantially to shifts in primary productivity throughout the open waters of the Arctic Ocean. However, it is clear that the response of primary production to sea ice loss has been both seasonally and spatially dependent (e.g., Tremblay et al. 2015). Furthermore, massive under-ice phytoplankton blooms have also recently been observed in Arctic waters (e.g., Arrigo et al. 2012, Arrigo et al. 2014). It is not clear whether these under-ice phytoplankton blooms are new phenomena. However, a shift away from snow-covered multi-year ice (typical of these areas in the 1980s) towards a thinner, more melt-ponded sea ice cover (typical of current conditions) will enhance the light transmittance (Frey et al., 2011) necessary for primary production, given the presence of sufficient nutrients.

In addition to phytoplankton primary production, sea ice algal production is also important to consider in the overall Arctic Ocean system. In the central Arctic Ocean where primary productivity is relatively low, sea ice algae can contribute up to 60% of total primary production (owing primarily to low pelagic primary productivity). Further, sea ice algae have recently been found to be principally limited by nitrate off the slope from the Laptev Sea and silicate at the ice margin near the Atlantic inflow (Fernández-Méndez et al. 2015). Important remaining questions include whether the production of sea ice algae in the central Arctic Ocean has increased over recent years owing to thinning ice and if nutrients are sufficient to sustain an increase in phytoplankton concentration with overall sea ice retreat (e.g., Fernández-Méndez et al. 2015).

Chlorophyll-*a*

Algae are responsible for nearly all photosynthesis occurring in the oceans. Measurements of the algal pigment chlorophyll (e.g. chlorophyll-*a*) serve as a proxy for the amount of algal biomass present as well as overall plant health. Here we present the complete, updated MODIS-Aqua satellite chlorophyll-*a* record for 2003-2016. A base period of 2003-2015 was chosen when calculating the 2016 anomalies to maximize the length of the short satellite-based time series. Since it is not possible to utilize satellite imagery to assess the presence of under-ice phytoplankton or sea ice algae, we focus here on satellite-based observations of chlorophyll-*a* concentrations and primary productivity in open ocean waters.

The 2016 data show anomalously high chlorophyll-*a* concentrations in a number of locations across the Arctic Ocean region, where patterns are spatially and temporally heterogeneous (**Fig. 6.1**). The most notable anomaly in 2016 occurred during May 2016, with high concentrations of chlorophyll-*a* (averaging over ~14 mg m^{-3} higher than the 2003-2015 mean) occurring across a relatively large (~300 x 300 km) region in the central Barents Sea, (**Figs. 6.1a, ee**) and possibly associated with earlier breakup of sea ice in that region (**Fig. 6.1i**). Weaker negative anomalies of chlorophyll-*a* (-2 to -5 mg m^{-3}) during May, June, July, and August were found in the Laptev Sea to the north and west of the New Siberian Islands (**Figs. 6.1e-h**), and may be a response to a slightly different trajectory of sea ice breakup (i.e., sea ice initially opening up to the northeast of the New Siberian Islands, rather than the northwest as has been more common over the previous decade (**Fig. 6.1k**)). One additional anomaly of note occurred in the Greenland Sea northeast of Greenland during June (**Fig. 6.1f**) over a ~150 x 750 km band along the sea ice edge in that region (**Fig. 6.1j**) and averaged ~7 mg m^{-3} higher than the 2003-2015 mean.

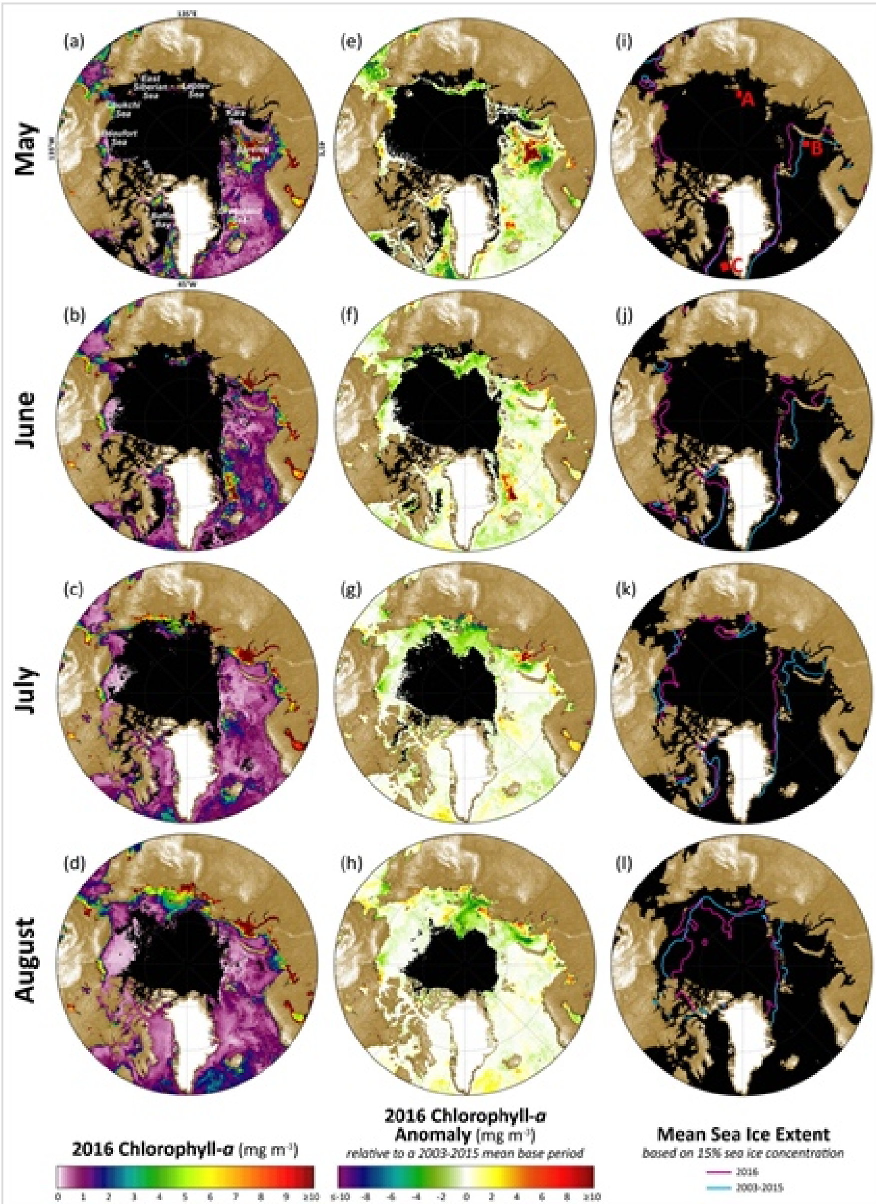

Fig. 6.1. Satellite-based chlorophyll-*a* data across the pan-Arctic region derived using the MODIS-Aqua Reprocessing 2014.0, OC3 algorithm: http://oceancolor.gsfc.nasa.gov/. Chlorophyll-*a* concentrations are shown here to foster direct measurements of ocean color and minimize the use of model output. Mean monthly chlorophyll-*a* concentrations during 2016 are shown for (a) May, (b) June, (c) July and (d) August. (e-h) Monthly anomalies of chlorophyll-*a* concentrations for 2016 (relative to a 2003-2015 mean base period). Black areas (a-h) denote a lack of data owing to either clouds or sea ice. (i-l) Sea ice extent (designated by a 15% sea ice concentration threshold) based on SSM/I data (Cavalieri et al., 1996; Maslanik and Stroeve, 1999) for the 2003-2015 mean (cyan line) and 2016 (magenta line). The locations A, B and C in (i) indicate the locations for the time series shown in **Fig. 6.3.**

The most significant rates of change in the 2003-2016 satellite record in May have occurred southwest of Greenland in the Labrador Sea and to the west of Novaya Zemlya in the Barents Sea (**Fig. 6.2a**).The latter has been associated with declines in sea ice (**Fig. 6.1i**) linked to the Atlantic Water inflow (e.g., Alexeev et al. 2013). Trends have been less steep during June overall (**Fig. 6.2b**), although some increases were also found in the Barents Sea northwest of Novaya Zemlya and northwest of Svalbard in the Greenland Sea. During July and August, significant increasing trends in chlorophyll-*a* concentrations have been found primarily in the Laptev Sea to the northwest of the New Siberian Islands (**Figs. 6.2c, d**). These particular trends are also linked to declining sea ice cover (**Figs. 6.1k, l**).

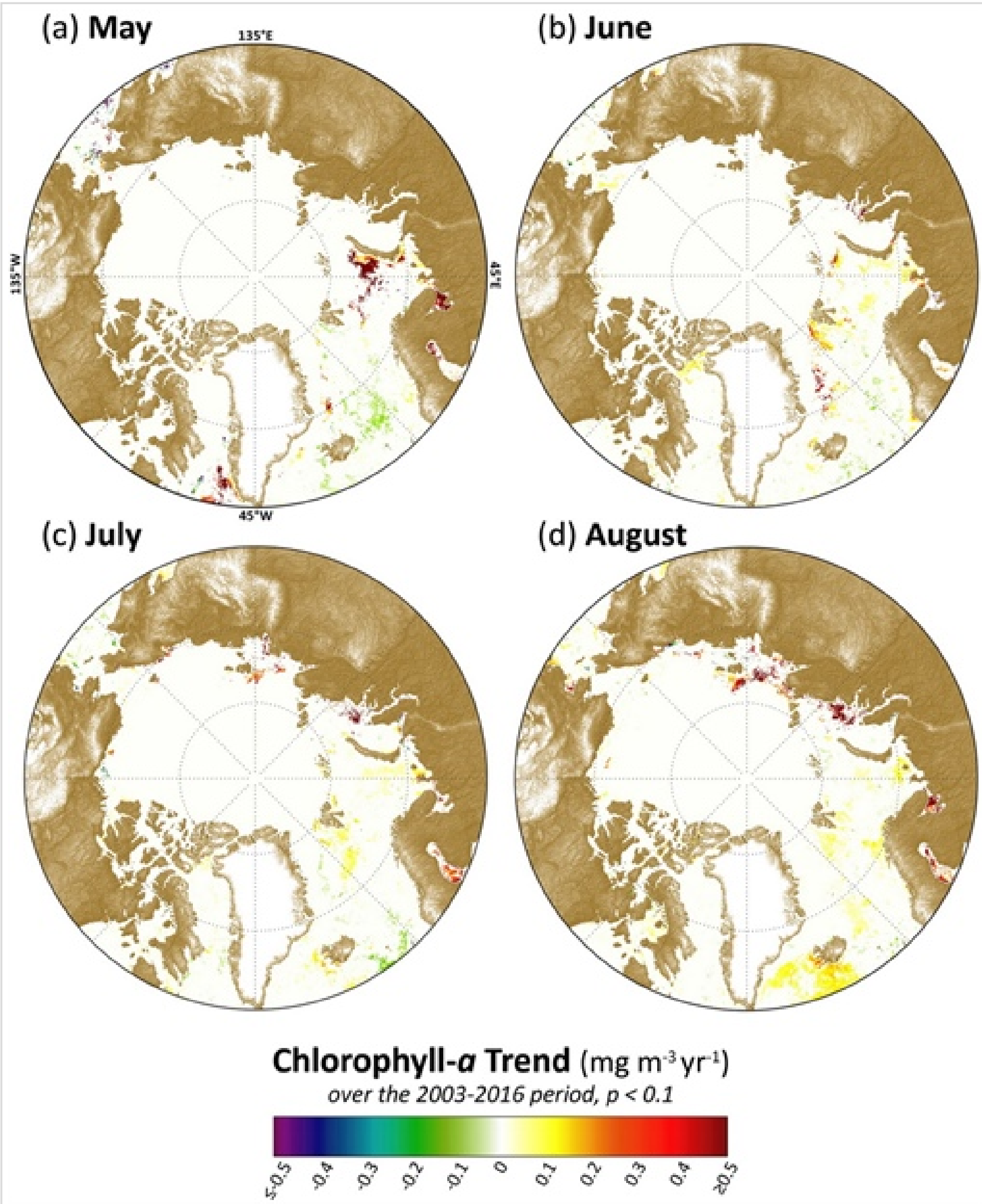

Fig. 6.2. Linear trends in satellite-based chlorophyll-*a* data across the pan-Arctic region derived using the MODIS-Aqua Reprocessing 2014.0, OC3 algorithm: http://oceancolor.gsfc.nasa.gov/. Theil-Sen median trends (2003-2016) in chlorophyll-*a* concentrations for each of the four months are shown, highlighting only statistically significant trends ($p < 0.1$, using the Mann-Kendall test for trend). Only those pixels with no more than 29% missing data (i.e., no more than 4 of 14 years of missing data) were included in the trend calculations, to maintain a robust statistic (Hoaglin et al. 2000).

To illustrate the quantitative nature of these trends, three example "hotspot" regions (shown in **Fig. 6.1i**) with notably steep trends in chlorophyll-*a* for May, June, July and August 2003-2016 are shown in **Fig. 6.3**. The region in the Laptev Sea (A), northwest of the New Siberian Islands, shows significant (p < 0.1) trends during July and August, while regions in the Barents Sea (B), west of Novaya Zemlya, and the Labrador Sea (C), southwest of Greenland, show significant (p < 0.1) trends during May. Of these trends, May 2016 made the largest contribution towards continued positive trends in chlorophyll-*a* at the Barents Sea site (**Fig. 6.3b**), with an overall positive trend of ~0.79 mg m^{-3} yr^{-1}. Even though trends during July and August at the Laptev Sea site (**Fig. 6.3a**) and during May at the Labrador Sea site (**Fig. 6.3c**) remain positive, concentrations at these two sites decreased considerably in 2016 compared to 2014 and 2015. In particular, the Laptev and Labrador Sea sites show high inter-annual variability in chlorophyll-*a* concentrations, with a rapid decrease in August and May 2016, respectively. In contrast, the Barents Sea site has exhibited a more gradual, secular trend during May over the 2003-2016 time series.

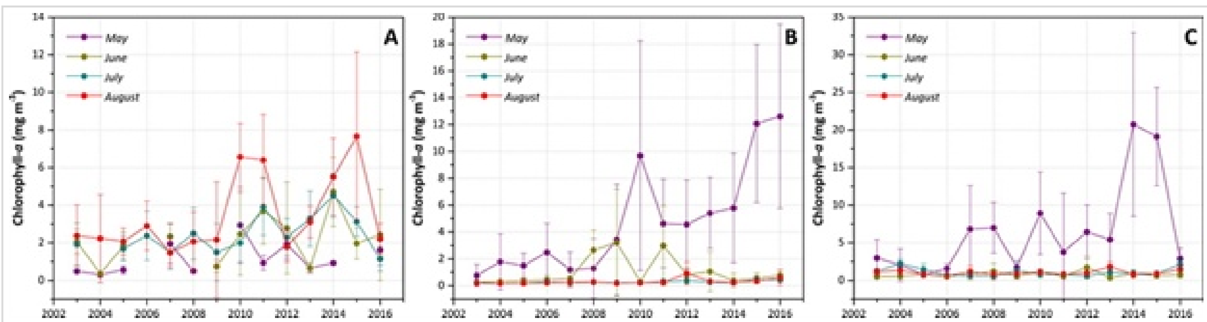

Fig. 6.3. Mean (±1 standard deviation) monthly chlorophyll-*a* concentrations (based on MODIS-Aqua satellite data) for May, June, July and August 2003-2016 for three example "hotspot" locations with notably steep trends in chlorophyll-*a*. The locations (shown in **Fig. 1i**) include (A) a ~22,500 km^2 region in the Laptev Sea northwest of the New Siberian Islands; (B) a ~25,400 km^2 region in the Barents Sea west of Novaya Zemlya; and (C) a 32,600 km^2 region in the Labrador Sea southwest of Greenland.

Primary Production

Estimates of ocean primary productivity for nine regions (and the average of these nine regions) across the Arctic show positive trends during the period 2003-2016 in all regions, except for the western (North America) Arctic (**Fig. 6.4**, **Table 6.1**). Similarly, anomalies in primary production for 2016 are positive for all regions except for the western Arctic and Sea of Okhotsk (**Table 6.1**). Regions with the highest anomalies for 2016 include the Barents Sea, the Greenland Sea, and Baffin Bay/Labrador Sea. Statistically significant trends between 2003 and 2016 occurred in the eastern (Eurasian) Arctic, Barents Sea, Greenland Sea, Hudson Bay and North Atlantic, with the steepest trends found for the Barents Sea and the eastern Arctic. There were no statistically significant trends for the western Arctic, Sea of Okhotsk, Bering Sea, or Baffin Bay/Labrador Sea.

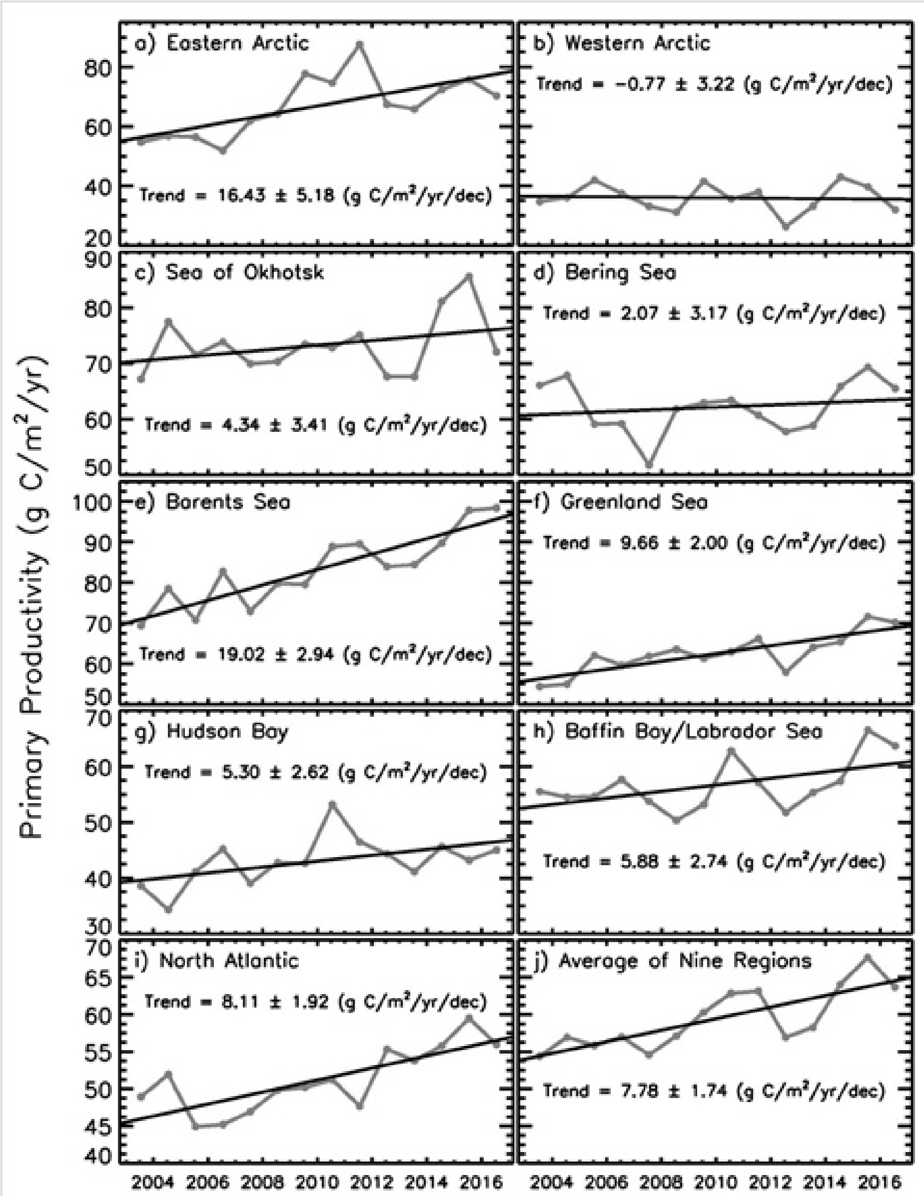

Fig. 6.4. Primary productivity (2003-2016, March-September) in nine different regions of the Northern Hemisphere, as well as the average of these nine regions, derived using chlorophyll-*a* concentrations from MODIS-Aqua data, AVHRR sea surface temperature data, and additional parameters. Values are calculated based on the techniques described by Behrensfield and Falkowski (1997) and represent net primary productivity (NPP). The trend and the statistical error of the slope (as derived from the regression analysis) are also shown for each region. Regions defined per Comiso (2015). Additional information regarding these data can be found in **Table 6.1.**

Table 6.1. Linear trends, statistical significance, percent change and primary productivity anomalies in 2016 (March-September) in the nine regions, and overall average, as shown in **Fig. 6.4**. Utilizing the Mann-Kendall test for trend, values in **bold** are significant at the 95% confidence level. The percent change is estimated from the linear regression of the 14-year time series.

Region	Trend, 2003-2016 (g C/m^2/yr/dec)	Mann-Kendall *p*-value	% Change	2016 Anomaly (g C/m^2/yr) *from a 2003-2015 base period*	2016 Anomaly (%) *from a 2003-2015 base period*
Eastern Arctic	**16.43**	0.010	37.9	3.58	5.4
Western Arctic	-0.77	0.914	-2.7	-4.33	-11.9
Sea of Okhotsk	4.34	0.451	8.8	-1.29	-1.8
Bering Sea	2.07	0.667	4.4	3.61	5.8
Barents Sea	**19.02**	0.000	34.8	16.25	19.8
Greenland Sea	**9.66**	0.001	22.3	8.15	13.1
Hudson Bay	**5.30**	0.047	17.4	2.14	5.0
Baffin Bay/Labrador Sea	5.88	0.193	14.5	7.50	13.3
North Atlantic	**8.11**	0.001	23.0	5.11	10.0
Average of Nine Regions	**7.78**	0.000	18.6	4.52	7.6

While similar trends have been reported previously for these regions using both SeaWiFS and MODIS data (Comiso 2015), satellite evidence does suggest that recent increases in cloudiness have dampened the increases in production that would have otherwise occurred as a function of sea ice decline alone (Bélanger et al. 2013). Further challenges remain with linking primary productivity, as well depth-integrated chlorophyll biomass throughout the water column, to satellite-based surface chlorophyll-*a* values (Tremblay et al. 2015). Satellite-based chlorophyll-*a* and primary productivity estimates continue to be confounded by issues such as river turbidity in coastal regions (e.g., Demidov et al. 2014, Chaves et al. 2015). Efforts to improve satellite retrieval algorithms based on *in situ* observations are thus critical to continue in all regions of the Arctic.

References

Alexeev, V. A., V. V. Ivanov, R. Kwok, and L. H. Smedsrud, 2013: North Atlantic warming and declining volume of arctic sea ice. *The Cryosphere Discuss.*, 7, 245-265, doi:10.5194/tcd-7-245-2013.

Arrigo, K. R., D. K. Perovich, R. S. Pickart, Z. W. Brown, G. L. van Dijken, K. E. Lowry, M. M. Mills, M. A. Palmer, W. M. Balch, F. Bahr, N. R. Bates, C. Benitez-Nelson, B. Bowler, E. Brownlee, J. K. Ehn, K. E. Frey, R. Garley, S. R. Laney, L. Lubelczyk, J. Mathis, A. Matsuoka, B. G. Mitchell, G. W. K. Moore, E. Ortega-Retuerta, S. Pal, C. M. Polashenski, R. A. Reynolds, B. Scheiber, H. M. Sosik, M. Stephens & J. H. Swift, 2012: Massive phytoplankton blooms under Arctic sea ice. *Science* 336, 1408, doi:10.1126/science.1215065.

Arrigo, K. R., D. K. Perovich, R. S. Pickart, Z. W. Brown, G. L. van Dijken, K. E. Lowry, M. M. Mills, M. A. Palmer, W. M. Balch, N. R. Bates, C. R. Benitez-Nelson, E. Brownlee, K. E. Frey, S. R. Laney, J. Mathis, A. Matsuoka, B. G. Mitchell, G. W. K. Moore, R. A. Reynolds, H. M. Sosik & J. H. Swift, 2014: Phytoplankton blooms beneath sea ice in the Chukchi Sea. *Deep Sea Research II*, http://dx.doi.org/10.1016/j.dsr2.2014.03.018.

Barber, D. G., H. Hop, C. J. Mundy, B. Else, I. A. Dmitrenko, J. -É. Tremblay, J. K. Ehn, P. Assmy, M. Daase, L. M. Candlish, and S. Rysgaard. 2015: Selected physical, biological and biogeochemical implications of a rapidly changing Arctic Marginal Ice Zone. *Progress in Oceanography*, doi:10.1016/j.pocean.2015.09.003.

Behrenfeld, M. J., and P. G. Falkowski, 1997: Photosynthetic rates derived from satellite-based chlorophyll concentration. *Limnology and Oceanography*, 42(1), 1-20.

Bélanger, S., M. Babin, and J. É. Tremblay, 2013: Increasing cloudiness in Arctic damps the increase in phytoplankton primary production due to sea ice receding. *Biogeosci*., 10, 4087-4101, doi:10.5194/bg-10-4087-2013.

Cavalieri, D. J., C. L. Parkinson, P. Gloersen, and H. Zwally, 1996, updated yearly: *Sea Ice Concentrations from Nimbus-7 SMMR and DMSP SSM/I-SSMIS Passive Microwave Data*. [2003-2014]. Boulder, Colorado USA: NASA DAAC at the National Snow and Ice Data Center.

Chaves, J., P. J. Werdell, C. W. Proctor, A. R. Neeley, S. A. Freeman, C. S. Thomas, and S. B. Hooker, 2015: Assessment of ocean color data records from MODIS-Aqua in the western Arctic Ocean. *Deep-Sea Research II*, 118, Part A, 32-43, doi:10.1016/j.dsr2.2015.02.011.

Comiso, J. C., 2015: Variability and trends of the Global Sea Ice Covers and Sea Levels: Effects on Physicochemical Parameters. *Climate and Fresh Water Toxins*, Luis M. Botana, M. Carmen Lauzao and Natalia Vilarino, Eds., De Gruyter, Berlin, Germany.

Demidov, A. B., S. A. Mosharov, and P. N. Makkaveev, 2014: Patterns of the Kara Sea primary production in autumn: Biotic and abiotic forcing of subsurface layer. *J. Marine Systems*, 132, 130-149, doi:10.1016/j.jmarsys.2014.01.014.

Fernandez-Mendez, M., C. Katlein, B. Rabe, M. Nicolaus, I. Peeken, K. Bakker, H. Flores, and A. Boetius, 2015: Photosynthetic production in the central Arctic Ocean during the record sea-ice minimum in 2012. *Biogeosci*., 12, 3525-3549.

Frey, K. E., D. K. Perovich, and B. Light, 2011: The spatial distribution of solar radiation under a melting Arctic sea ice cover. *Geophysical Research Letter*s 38, L22501, doi:10.1029/2011GL049421.

Hoaglin, D. C., F. Mosteller, and J. W. Tukey, 2000: *Understanding Robust and Exploratory Data Analysis*, Wiley Classics Library Edition, (New York: Wiley).

Leu, E., C. J. Mundy, P. Assmy, K. Campbell, T. M. Gabrielsen, M. Gosselin, T. Juul-Pedersen, and R. Gradinger, 2015: Arctic spring awakening - Steering principles behind the phenology of vernal ice algal blooms. *Progress in Oceanography*, http://dx.doi.org/10.1016/j.pocean.2015.07.012.

Maslanik, J., and J. Stroeve, 1999, updated daily: *Near-Real-Time DMSP SSM/I-SSMIS Daily Polar Gridded Sea Ice Concentrations*. [2016]. Boulder, Colorado USA: NASA DAAC at the National Snow and Ice Data Center.

Müller-Karger, F. E., R. Varela, R. Thunell, R. Luerssen, C. Hu, and J. J. Walsh, 2005: The importance of continental margins in the global carbon cycle. *Geophys. Res. Lett.*, 32, L01602, doi:10.1029/2004GL021346.

Tremblay J. -É., L. G. Anderson, P. Matrai, S. Bélanger, C. Michel, P. Coupel, and M. Reigstad, 2015: Global and regional drivers of nutrient supply, primary production and CO_2 drawdown in the changing Arctic Ocean. *Progress in Oceanography*, doi:10.1016/j.pocean.2015.08.009.

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Tundra Greenness

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Highlights

- Tundra greenness increased throughout the Arctic in 2015 (relative to 2014), following 3-4 years of continuous declines; an exception was that time-integrated (TI)-NDVI (an indicator of vegetation productivity) continued to decrease in North America.
- Long-term trends (1982-2015) show greening on the North Slope of Alaska, in the southern Canadian tundra, and in much of the central and eastern Siberian tundra, and browning in western Alaska (Yukon-Kuskokwim Delta), the higher-Arctic Canadian Archipelago, and western Siberian tundra.

Vegetation in the Arctic tundra has been responding dynamically over the course of the last several decades to environmental changes, many of which are anthropogenically-induced. These vegetation changes throughout the circumpolar Arctic are not spatially homogeneous, nor are they temporally consistent (e.g. Bhatt et al. 2013), suggesting that there are complex interactions among atmosphere, ground (soils and permafrost), vegetation, and herbivore components of the Arctic system. Changes in Arctic tundra vegetation may have a relatively small impact on the global carbon budget through photosynthetic uptake of CO_2, compared to changes in other carbon cycling processes (Abbott et al. 2016). However, tundra vegetation can have important effects on permafrost, hydrological dynamics, soil carbon fluxes, and the surface energy balance (e.g. Blok et el. 2010, Myers-Smith and Hik 2013, Parker et al. 2015). Tundra vegetation dynamics also control the diversity of herbivores (birds and mammals) in the Arctic, with species richness being positively related to vegetation productivity (Barrio et al. 2016). To improve our understanding of these complex interactions and their impacts on the Arctic and global systems, we continue to evaluate the state of the circumpolar Arctic vegetation.

Using Earth-Observing Satellites (EOS) with daily return intervals, we have had the capacity to monitor Arctic tundra vegetation continuously since 1982. We report on data from the Global Inventory Modeling and Mapping Studies (GIMMS) version 3g dataset (GIMMS 2013) based largely on the Advanced Very High Resolution Radiometer (AVHRR) sensor onboard NOAA satellites (Pinzon and Tucker 2014). The GIMMS product is a bi-weekly, maximum-value composited dataset of the Normalized Difference Vegetation Index (NDVI); NDVI is highly correlated with aboveground vegetation (e.g. Raynolds et al. 2012), or tundra "greenness." We use two metrics based on the NDVI: MaxNDVI and TI-NDVI. Max NDVI is the peak NDVI for the year (growing season), and is related to yearly maximum aboveground vegetation biomass. TI (time-integrated) NDVI is the sum of the bi-weekly NDVI values for the growing season, and is related to the total aboveground vegetation productivity.

Examining the overall trend in tundra greenness for the 34-year record, we find that both MaxNDVI and TI-NDVI have increased on the North Slope of Alaska, in the southern Canadian tundra, and in much of the central and eastern Siberian tundra, whereas tundra greenness has decreased (i.e. "Arctic browning") in western Alaska (Yukon-Kuskokwim Delta), the higher-Arctic Canadian Archipelago, and western Siberian tundra (**Fig. 7.1**). Using the same NDVI dataset (albeit with a different vegetation map and a slightly shorter duration–1982-2012), Loranty et al. (2016) found that a much greater fraction of tundra areas overlying continuous permafrost exhibited long-term greening (42%) compared to browning (5%); in tundra areas overlying discontinuous permafrost the areal difference was not as great (27% greening and 10% browning). Across Arctic vegetation types (from 1982-2014), greening has been most extensive in forest - tall shrub tundra, moderately extensive in shrub tundra and sedge tundra, and minimal in low-lying shrub tundra. Interestingly, forest - tall shrub tundra also had the greatest fractional area of browning among the vegetation types, although the area of browning was <8% that of the area greening (Park et al. 2016). If we assess the 34-year trends with more temporal detail, both the North American and Eurasian Arctic have shown substantial increases in tundra greenness up to the early 2010s for MaxNDVI and the early 2000s for TI-NDVI. Since then, declines are visible in these tundra greenness indices (**Fig. 7.2**).

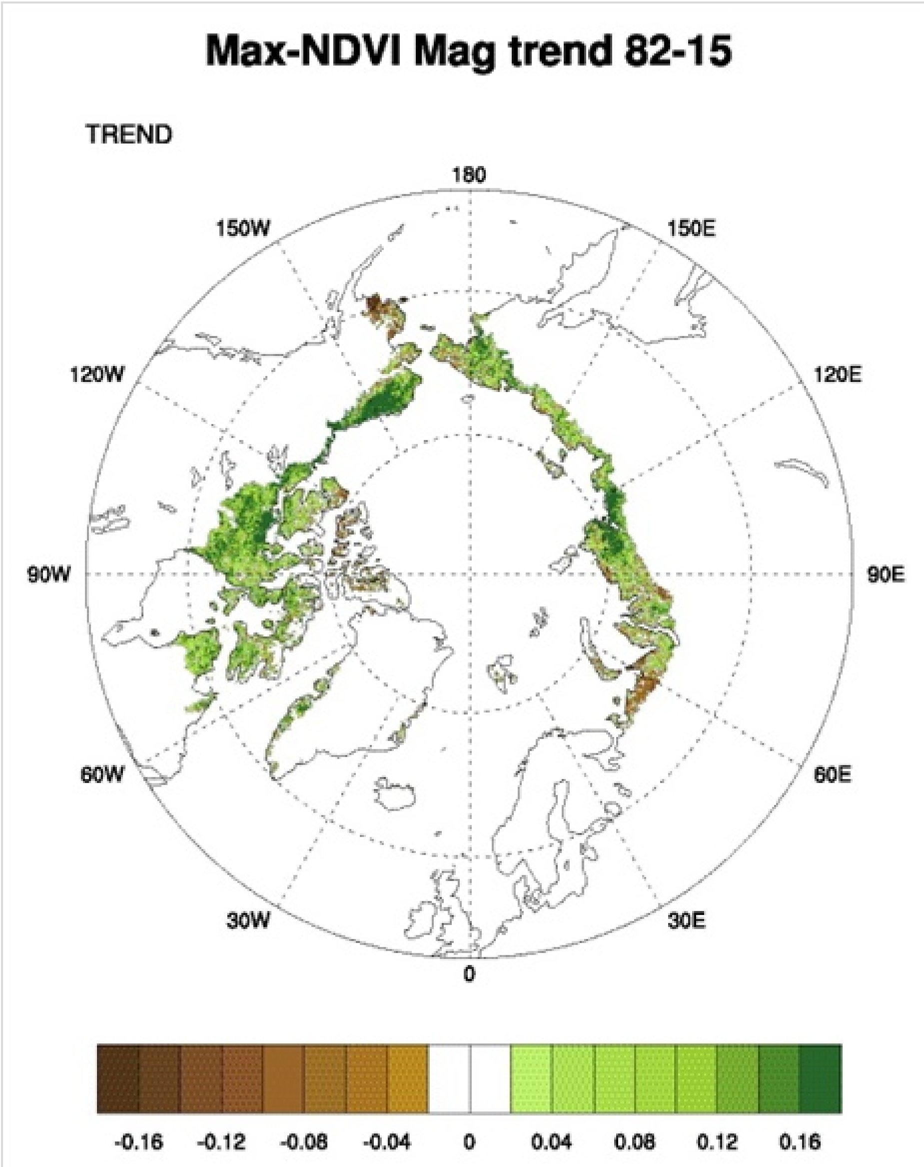

Fig. 7.1a Magnitude of the trend in maximum (Max)-NDVI from 1982-2015.

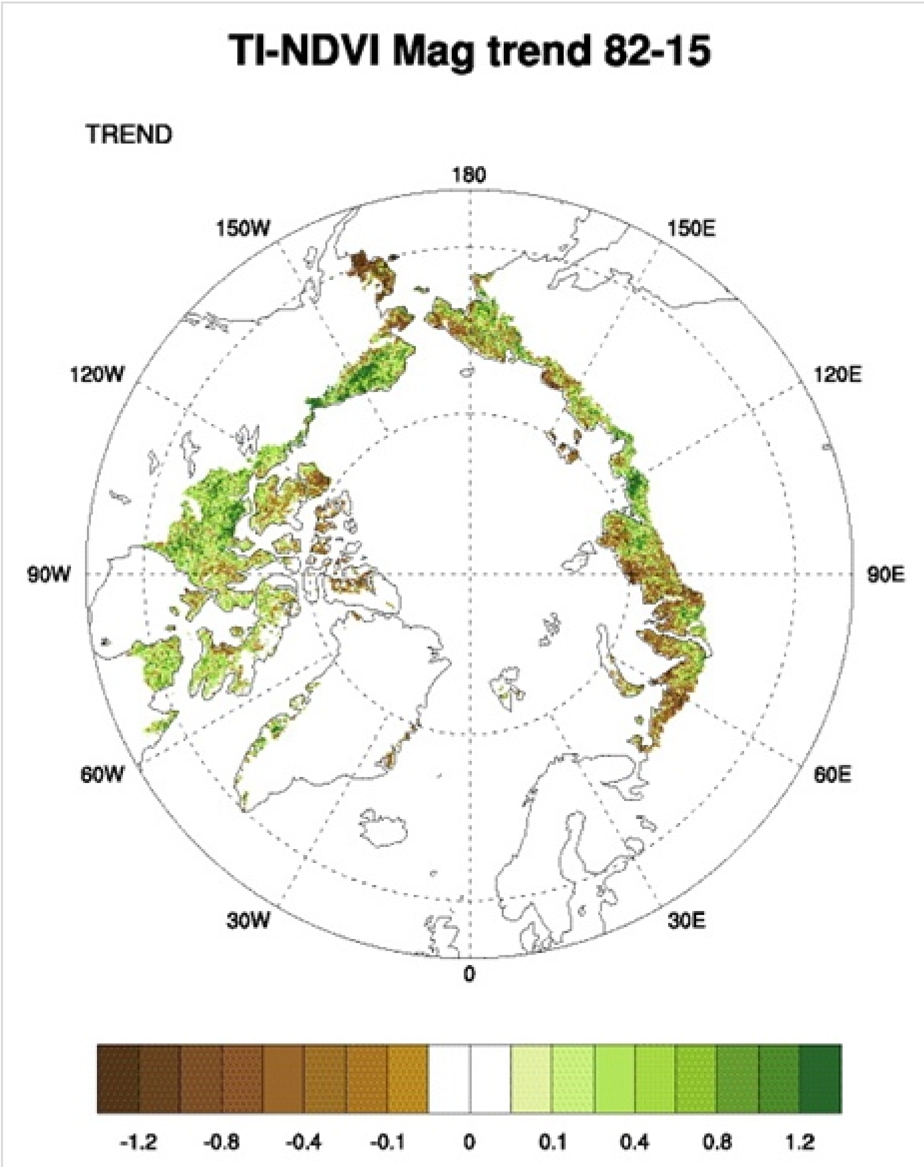

Fig. 7.1b. Magnitude of the trend in temporal-integrated (TI)-NDVI from 1982-2015.

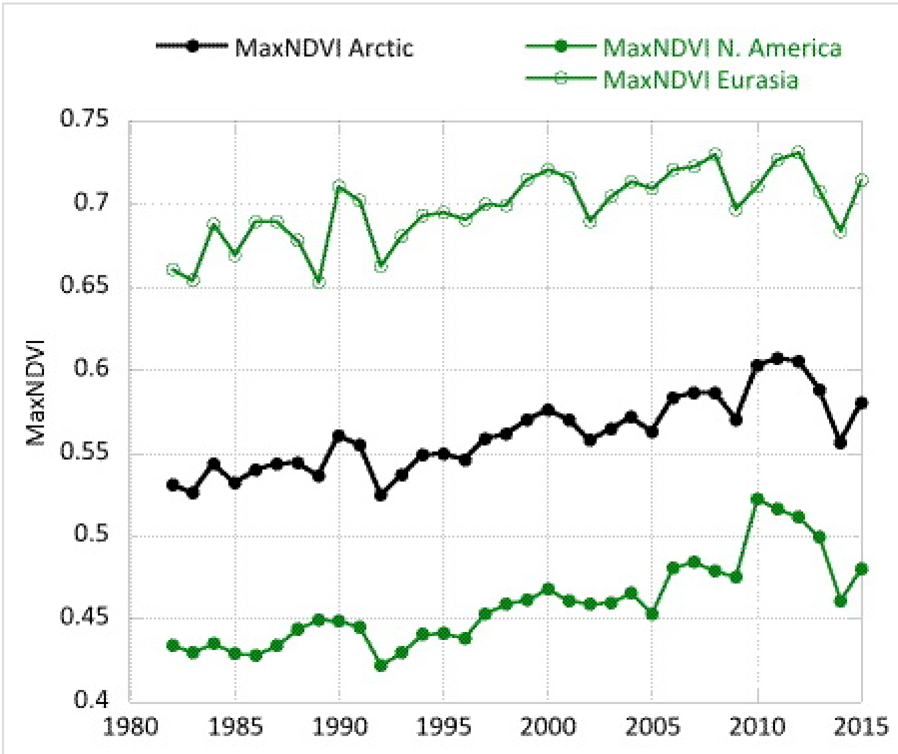

Fig. 7.2a. Maximum (Max) NDVI from 1982 to 2015 for North America (bottom), Eurasia (top), and the Arctic as a whole (middle).

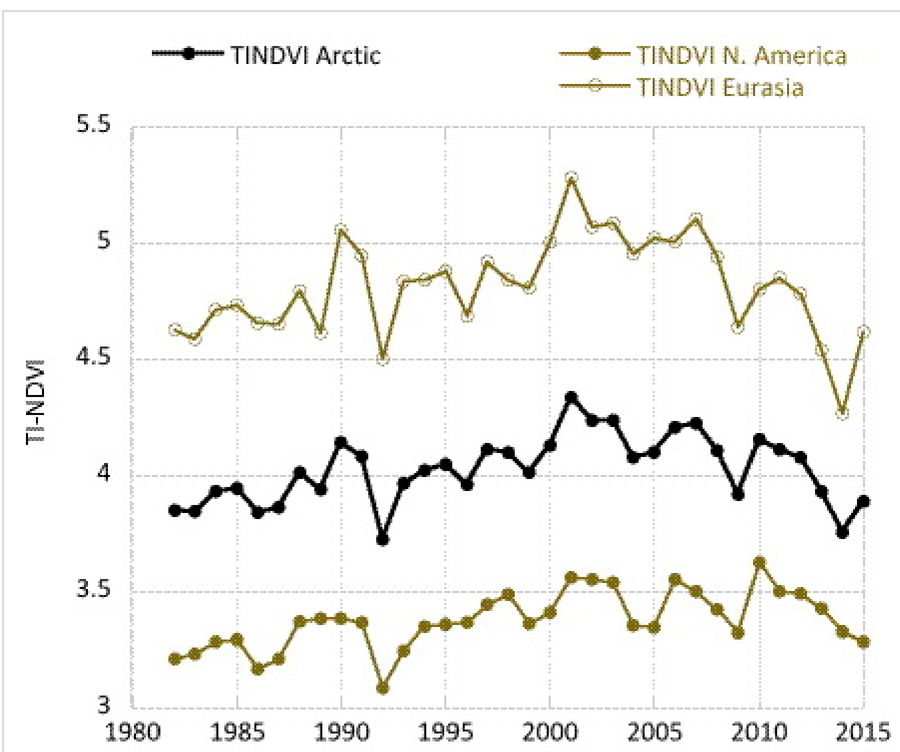

Fig. 7.2b. Time-integrated (TI)-NDVI from 1982 to 2015 for North America (bottom), Eurasia (top), and the Arctic as a whole (middle).

Following 3-4 years of successive declines (depending on the index and the continent), the NDVI for Arctic tundra exhibited an upturn during the summer of 2015, with the exception of TI-NDVI for North America, which continued to decrease (**Fig. 7.2**). Based on remotely-sensed Land Surface Temperatures (LST), from the same sensors as those providing the NDVI values, cumulative summer warmth (sum of mean monthly temperatures > 0° C) for the Arctic as a whole (and for the two continents separately) was greater in 2015 than in any other year of the satellite record (since 1982). Correspondingly, MaxNDVI increased by 4.3% and TI-NDVI increased by 3.6% for the circumpolar Arctic in 2015, relative to 2014 values. MaxNDVI values in 2015 were greater than the mean values for the record (1982-2015), ranking 8, 7, and 9 for the Arctic, North American Arctic, and Eurasian Arctic respectively over the 34-year record. TI-NDVI values were below the mean for the entire record, ranking 28, 28, and 29 for the Arctic, North American Arctic, and the Eurasian Arctic respectively.

While research on tundra browning is at present relatively sparse, there may be a variety of mechanisms leading to browning. Tundra browning in certain Arctic regions has been attributed to cooler summer temperatures (Bhatt et al. 2013) and to deeper winter snow packs and potentially longer snow cover duration (Bieniek et al. 2015). Again using the same NDVI dataset, Park et al. (2016) noted that since 2000 there has been a shortening of the growing season in the northern high latitudes, due to both a delayed start and an earlier end to the season. In a recent Commentary, Phoenix and Bjerke (2016) propose that tundra browning could be more "event-driven" than greening, caused by fire (Bret-Harte et al. 2013), extreme winter warming (Bokhorst et al. 2011), other anomalous weather events (e.g. frost damage), as well as outbreaks of insect and fungal pests (Graglia et al. 2001, Bjerke et al. 2014). Another potential cause of tundra browning could be increases in herbivore populations (Pederson et al. 2013, Hupp et al. 2015, Barrio et al. 2016).

In a recent remote-sensing analysis of global terrestrial ecosystems, Seddon et al. (2016) indicated that the Arctic tundra has been highly sensitive to climate variability over the past 14 years, the length of the satellite-based Moderate Resolution Imaging Spectroradiometer (MODIS) record. They also suggest that this sensitivity is largely correlated with temperature and cloudiness, environmental variables that are presently being altered (and likely will continue to be altered) by anthropogenic climate change. Further, Seddon et al. (2016) report much greater vegetation sensitivity to climatic variability in the low and mid-latitude tundra regions than in the High Arctic, in agreement with other remote sensing results (Epstein et al. 2012) and those of Myers-Smith et al. (2015), based on *in situ* growth measurements. Fully understanding the mechanisms behind tundra vegetation dynamics (and subsequent effects on other system properties) will require further cooperative efforts in scientific monitoring (field and remote sensing), experimentation, and simulation modeling.

References

Abbott, B. W., J. B. Jones, E. A. G. Schuur, F. S. Chapin III, W. B. Bowden, M. S. Bret-Harte, H. E. Epstein, M. D. Flanagan, T. K. Harms, T. N. Hollingsworth, M. C. Mack, A. D. McGuire, S. M. Natali, A.V. Rocha, S. E. Tank, M. R. Turetsky, J. E. Vonk, K. P. Wickland, G. R. Aiken, H. D. Alexander, R. M. W. Amon, B. W. Benscoter, Y. Bergeron, K. Bishop, O. Blarquez, B. Bond-Lamberty, A. L. Breen, I. Buffam, Y. Cai, C. Carcaillet, S. K. Casey, J. M. Chen, H.

Y. H. Chen, T. R. Christensen, L. W. Cooper, J. H. C. Cornelissen, W. J. de Groot, T. H. DeLuca, E. Dorrepaal, N. Fletcher, J. C. Finlay, B. C. Forbes, N. H. F. French, S. Gauthier, M. P. Girardin, S. J. Goetz, J. G. Goldhammer, L. Gough, P. Grogan, L. Guo, P. E. Higeura, L. Hinzman, F. S. Hu, G. Hugelius, E. E. Jafarov, R. Jandt, J. F. Johnstone, I. Karlsson, E. S. Kasischke, G. Kattner, R. Kelly, F. Keuper, G. W. Kling, P. Kortelainen, J. Kouki, P. Kuhry, J. Laudon, I. Laurion, R. W. Macdonald, P. J. Mann, P. J. Martikainen, J. W. McClelland, U. Molau, S. F. Oberbauer, D. Olefeldt, D. Paré, M. -A. Parisien, S. Payette, C. Peng, O. S. Pokrovsky, E. B. Rastetter, P. A. Raymond, M. K. Raynolds, G. Rein, J. F. Reynolds, M. Robard, B. M. Rogers, C. Schädel, K. Schaefer, I. K. Schmidt, A. Shvidenko, J. Sky, R. G. M. Spencer, G. Starr, R. G. Striegl, R. Teisserenc, L. J. Tranvik, T. Virtanen, J. M. Welker, and S. Zimov, 2016: Biomass offsets little or none of permafrost carbon release from soils, streams, and wildfire: an expert assessment. *Environm. Res. Lett.*, 11, 034014.

Barrio, I. C., C. G. Bueno, M. Gartzi, E. M. Soininen, K. S. Christie, J. D. M. Speed, V. T. Ravolainen, B. C. Forbes, G. Gauthier, T. Horstkotte, K. S. Hoset, T. T. Høye, I. S. Jónsdóttir, E. Lévesque, M. A. Mörsdorf, J. Olofsson, P. A. Wookey, and D. S. Hik, 2016: Biotic interactions mediate patterns of herbivore diversity in the Arctic. *Glob. Ecol. Biogoegr.*, doi:10.1111/geb.12470.

Bhatt, U. S., D. A. Walker, M. K. Raynolds, P. A. Bieniek, H. E. Epstein, J. C. Comiso, J. E. Pinzon, C. J. Tucker, and I. V. Polyakov, 2013: *Recent declines in warming and arctic vegetation greening trends over pan-Arctic tundra, Remote Sens*. (Special NDVI3g Issue), 5, 4229-4254; doi:10.3390/rs5094229.

Bieniek, P. A., U. S. Bhatt, D. A. Walker, M. K. Raynolds, J. C. Comiso, H. E. Epstein, J. E. Pinzon, C. J. Tucker, R. L. Thoman, H. Tran, N. Mölders, M. Steele, J. Zhang, and W. Ermold, 2015: Climate drivers linked to changing seasonality of Alaska coastal tundra vegetation productivity. *Earth Interactions*, 19, 19.

Bjerke, J. W., S. R. Karlsen, K. A. Hogda, E. Malnes, J. U. Jepsen, S. Lovibond, D. Vikhamar-Schuler, and H. Tommervik, 2014: Record-low primary productivity and high plant damage in the Nordic Arctic Region in 2012 caused by multiple weather events and pest outbreaks. *Environm. Res. Lett.*, 9, 084006.

Blok, D., M. Heijmans, G. Schaepman-Strub, A. Kononov, T. Maximov, and F. Berendse, 2010: Shrub expansion may reduce summer permafrost thaw in Siberian tundra. *Glob. Change. Biol.*, 16, 1296-305.

Bokhorst, S., J. W. Bjerke, L. E. Street, T. V. Callaghan, and G. K. Phoenix, 2011: Impacts of multiple extreme winter warming events on sub-Arctic heathland: phenology, reproduction, growth, and CO_2 flux responses. *Glob. Change. Biol.*, 17, 2817-2830.

Bret-Harte, M. S., M. C. Mack, G. R. Shaver, D. C. Huebner, M. Johnston, C. A. Mojica, C. Pizano, and J. A. Reskind, 2013: The response of Arctic vegetation and soils following an unusually severe tundra fire. *Phil. Trans. Royal Soc. B-Biol. Sci.*, 368, doi:10.1098/rstb.2012.0490.

Epstein, H. E., M. K. Raynolds, D. A. Walker, U. S. Bhatt, C. J. Tucker, and J. E. Pinzon, 2012: Dynamics of aboveground phytomass of the circumpolar arctic tundra over the past three decades. *Environm. Res. Lett.* 7, 015506.

Global Inventory Modeling and Mapping Studies (GIMMS), 2013: Available online: http://gcmd.nasa.gov/records/GCMD_GLCF_GIMMS.html.

Graglia, E., R. Julkunen-Titto, G. R. Shaver, I. K. Schmidt, S. Jonasson, and A. Michelsen, 2001: Environmental controls and intersite variations of phenolics in Betula nana in tundra ecosystems. *New Phytologist*, doi:10.1046/j.1469-8137.2001.00149.x.

Hupp, J. W., D. H. Ward, M. E. Whalen, and J. M Pearce, 2015: Changing arctic ecosystems - What is causing the rapid increase of Snow Geese in northern Alaska? *U.S. Geological Survey* Fact Sheet 2015-3062, http://dx.doi.org/10.3133/fs20153062.

Loranty, M. M., W. Liberman-Cribbin, L. T. Berner, S. M. Natali, S. J. Goetz, H. D. Alexander, and A. L. Kholodov, 2016: Spatial variation in vegetation productivity trends, fire disturbance, and soil carbon across arctic-boreal permafrost ecosystems. *Environm. Res. Lett.*, 11, 095008.

Myers-Smith, I. H., S. C. Elmendorf, P. S. A. Beck, M. Wilmking, M. Hallinger, D. Blok, K. D. Tape, S. A. Rayback, M. Macias-Fauria, B. C. Forbes, J. D. M. Speed, N. Boulangier-Lapointe, C. Rixen, E. Lévesque, N. Martin Schmidt, C. Baittinger, A. J. Trant, L. Hermanutz, L. Siegwart Collier, M. A. Dawes, T. C. Lantz, S. Weijers, R. Haalfdan Jørgensen, A. Buchwal, A. Buras, A. T. Naito, V. Ravolainen, G. Schaepman-Strub, J. A. Wheeler, S. Wipf, K. C. Guay, D. S. Hik, and M. Vellend, 2015: Climate sensitivity of shrub growth across the tundra biome. *Nature Climate Change*, doi:10.1038/NCLIMATE2697.

Myers-Smith, I. H., and D. S. Hik, 2013: Shrub canopies influence soil temperatures but not nutrient dynamics: an experimental test of tundra snow-shrub interactions. *Ecol. Evol.* 3, 3683-700.

Park, T., S. Ganguly, H. Tømmervik, E. S. Euskirchen, K. -A. Høgda, S. R. Karlsen, V. Brovkin, R. R. Nemani, and R. B. Myneni, 2016: Changes in growing season duration and productivity of northern vegetation inferred from long-term remote sensing data. *Environm. Res. Lett.*, 11, 084001.

Parker, T. C., J. -A. Subke, and P. A. Wookey, 2015: Rapid carbon turnover beneath shrub and tree vegetation is associated with low soil carbon stocks at a sub-arctic treeline. *Glob. Change Biol.*, 21, 2070-2081.

Pedersen, Å. Ø., J. D. M. Speed, and I. M. Tombre. 2013: Prevalence of pink-footed goose grubbing in the arctic tundra increases with population expansion. *Polar Biol.*, 36, 1569-1575.

Phoenix, G. K., and J. W. Bjerke, 2016: Arctic browning: extreme events and trends reversing arctic greening. *Glob. Change. Biol.*, doi:10.1111/gcb.13261.

Pinzon, J., and C. Tucker, 2014: A non-stationary 1981-2014 AVHRR NDVI3g time series. *Remote Sens.*, 6, 6929-6960, doi:10.3390/rs6086929.

Raynolds M. K., D. A. Walker, H. E. Epstein, J. E. Pinzon, and C. J. Tucker, 2012: A new estimate of tundra-biome phytomass from trans-Arctic field data and AVHRR NDVI. *Remote Sens. Lett.*, 3, 403-411.

Seddon, A. W. R., M. Macias-Fauria, P. R. Long, D. Benz, and K. J. Willis, 2016: Sensitivity of global terrestrial ecosystems to climate variability. *Nature* 531, doi:10.1038/nature16986.

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Ocean Acidification

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Highlights

- The waters of the Arctic Ocean are disproportionately prone to ocean acidification compared to the rest of the global ocean due to the intensity, duration and extent of natural and anthropogenic drivers.
- Over the next 2-3 decades, it is likely that ocean acidification will continue to intensify, especially over the shallow Arctic shelves. The rapid rate of this change is likely to have detrimental effects on ecosystems that are already under pressure from rising temperatures and other climate-driven stressors.

Over the past five years, ocean acidification (OA) has emerged as one of the most prominent issues in marine research. This is especially true given the newfound public understanding of the potential biological threat to marine calcifiers (e.g. clams, pteropods) and associated fisheries, and the associated human impacts for small communities that directly or indirectly rely on them (e.g., Mathis et al., 2015a; Frisch et al., 2015). Cooler water temperatures and unique physical processes (i.e. formation and melting of sea ice) make the waters of the Arctic Ocean disproportionately sensitive to OA when compared to the rest of the global ocean. Even small amounts of human-derived carbon dioxide (CO_2) can cause significant chemical changes that other areas do not experience, and these could pose an existential threat to some biological organisms.

Current data indicates that certain areas of the Arctic shelves presently experience prolonged ocean acidification events in shallow bottom waters (Bates, 2015; Cross et al., 2016). Because natural carbon accumulation is already high near shelf sediments, these waters are most vulnerable to anthropogenic CO_2 accumulation. These waters are eventually transported off the shelf. As a result, new analysis released this year suggests that corrosive conditions have been expanding deeper into the Arctic Basin over the last several decades (Qi et al., 2016). These Pacific-origin corrosive waters have been observed as far as the entrances to Amundsen Gulf and M'Clure Strait in the Canadian Arctic Archipelago (Mathis et al., 2012; Cross et al., 2016). The formation and transport of corrosive waters on the Pacific Arctic shelves may have widespread impact on the Arctic biogeochemical system, compounding acidification all the way to the North Atlantic.

The inherently short Arctic food web linkages may lead to widespread impacts of OA on the Arctic marine ecosystem. For example, many upper-trophic level organisms (e.g. walrus, grey whales and salmon) rely on the marine calcifiers most likely to be impacted by OA (Cross et al., 2016). Juvenile and larval life stages of these upper trophic organisms are also particularly vulnerable to OA (e.g., crabs [Long et al., 2013a,b; Punt et al., 2014] and shellfish, [Ekstrom et al., 2015]). In turn, many subsistence communities rely on these upper trophic populations. While biological impacts of OA are not presently visible, it is likely that OA conditions will intensify over the next 2-3 decades and may produce more prominent impacts (Mathis et al., 2015b; Punt et al., 2016).

Present-day acidification hotspots are produced by a combination of natural CO_2 accumulation caused by biological cycles every season with additional, human-derived CO_2 absorbed from the atmosphere. Natural CO_2 accumulation tends to be highest in areas where currents are particularly slow, and where bottom waters are largely isolated from the surface. As anthropogenic CO_2 continues to increase, the Arctic Ocean will be pushed past important chemical thresholds even in areas where natural carbon accumulation is lower (Mathis et al., 2015b).

In the last twelve months, several comprehensive data synthesis products (Mathis et al., 2015b; Bates, 2015; Semiletov et al., 2016; Qi et al., 2016; Cross et al., 2016) have been published using much of the available OA data that has been collected in the Arctic Ocean. Several trends have emerged that clearly elucidate the rapid progression of OA across the Arctic basin, including rapid CO_2 uptake from the atmosphere and increasing carbonate mineral corrosivity (e.g., Evans et al., 2013). These trends are compounded by regional variability that is controlled by a number of physical and chemical processes including the rate of air-sea gas exchange of carbon dioxide (e.g. Bates, 2015; Evans et al., 2015); upwelling (Mathis et al., 2013); transport of allocthonous terrestrial carbon from river discharge and permafrost degradation (Semiletov et al., 2016); sea-ice melt (Yamamoto-Kawai et al., 2013); and respiration (Bates 2015).

Additionally, the indirect effect of changing sea-ice coverage is providing a positive feedback to OA (see essay on *Sea Ice*). For example, the reduction in Arctic and sub-Arctic sea ice observed in recent years can be attributed to increased warming caused at least in part by rising atmospheric CO_2 levels. The reduction in the extent and duration of sea-ice cover leads to longer open water periods, allowing for enhanced upwelling and changes in the timing and intensity of primary production (see essay on *Primary Productivity*) and sea-air gas exchange of CO_2 (e.g. Evans et al., 2015). Combined with the fact that cooler temperatures and global ocean circulation processes precondition the continental shelves in this region to have relatively low carbonate mineral saturation states compared to the rest of the ocean, these direct and indirect acidification processes make the region highly vulnerable to further reductions in seawater pH and saturation states.

Though the specifics remain uncertain, it is likely that the consequences of continuing OA will be detrimental for parts of the marine food web (Mathis et al., 2015a). Continued monitoring of OA in the Arctic Ocean is essential to understand the ecosystem transitions currently underway due to the suite of anthropogenic pressures. The Arctic region provides unique insights into how the global ocean will respond to human activities and it is our best hope for developing the understanding that will be needed to adapt to what will be our new, modern ocean environment.

References

Bates, N. R. 2015. Assessing ocean acidification variability in the Pacific-Arctic region as part of the Russian-American Long-term Census of the Arctic, *Oceanography* 28(3):36-45, http://dx.doi.org/10.5670/oceanog.2015.56.

Cross, J. N., Mathis, J. T., Pickart, R. S., Bates, N. R., 2016. Formation and transport of corrosive water in the Pacific Arctic region. *Deep-Sea Research II*, submitted.

Ekstrom, J. A., Suatoni, L., Cooley, S. R., Pendleton, L. H., Waldbusser, G. G., Cinner, J. E., Ritter, J., Langdon, C., van Hooidonk, R., Gledhill, D., Wellman, K., Bech, M. W., Brander, L. M., Rittschof, D., Doherty, C., Edwards, P. E. T., Portela, R., 2016. Vulnerability and adaptation of US shellfisheries to ocean acidification. *Nature Climate Change*, 5, 207-214, doi: 10.1038/nclimate2508.

Evans, W., Mathis, J. T., Cross, J. N., Bates, N. R., Frey, K. E., Else, B. G. T., Papakyriakou, T. N., DeGrandpre, M. D., Islam, F., Cai, W. -J., Chen, B., Yamamoto-Kawai, M., Carmack, E., Williams, W. J., Takahashi, T., 2015. Sea-air CO2 exchange in the western Arctic coastal ocean. *Glob. Biogeochem. Cycles*, 29(8), 1190-1209, doi: 10.1002/2015GB005153.

Frisch, L. C., J. T. Mathis, N. P. Kettle, and S. F. Trainor. 2015. Gauging perceptions of ocean acidification in Alaska. *Marine Policy* 53:101-110, http://dx.doi.org/10.1016/j.marpol.2014.11.022.

Long, W. C., Swiney, K. M., Harris, C., Page, H. N., Foy, R. J., 2013a. Effects of ocean acidification on juvenile Red King crab (Paralithodes camtschaticus) and Tanner crab (Chinoecetes bairdi) growth, condition, calcification, and survival. *PLOS ONE*, 8(4), e60959, doi:10.1371/journal.pone.0060959.

Long, W. C., Swiney, K. M., Foy, R. J., 2013b. Effects of ocean acidification on the embryos and larvae of red king crab, Paralithodes camtschaticus. *Mar. Pol. Bull.*, 69, 38-47, doi: 10.1016/j.marpolbul.2013.01.011.

Mathis, J. T., Pickart, R. S., Byrne, R. H., McNeil, C. L., Moore, G. W. K., Juranek, L. W., Liu, X., Ma, J., Easley, R. A., Elliot, M. M., Cross, J. N., Reisdorph, S. C., Bahr, F., Morison, J., Lichendorf, T., Feely, R. A., 2013. Storm-induced upwelling of high pCO2 waters onto the continental shelf of the western Arctic Ocean and implications for carbonate mineral saturation states. *Geophys. Res. Lett.*, 39, L07606, doi: 10.1029/2012GL051574.

Mathis, J. T., Cooley, S. R., Lucey, N., Colt, S., Ekstrom, J., Hurst, T., Hauri, C., Evans, W., Cross, J. N., Feely, R. A., 2015a. Ocean acidification risk assessment for Alaska's fishery sector. *Prog. Oceanogr.*, 136, 71-91, doi: 10.1016/j.pocean.2014.07.001.

Mathis, J. T., Cross, J. N., Doney, S. C., 2015b. Ocean acidification in the surface waters of the Pacific-Arctic boundary regions. *Oceanogr.*, 28(2), 122-135, doi: 1 0.5670/oceanog.2015.36.

Punt, A. E., Poljak, D., Dalton, M. G., Foy, R. J., 2014. Evaluating the impact of ocean acidification on fishery yields and profits: The example of red king crab in Bristol Bay. *Ecol. Modell.*, 2014, 285, 39-53, doi: 10.1016/j.ecolmodel.2014.04.017.

Punt, A. E., Foy, R. J., Dalton, M. G., Long, C., and Swiney, K. M., 2016. Effects of long-term exposure to ocean acidification conditions on future southern Tanner crab (Chionoecetes bairdi) fisheries management. *ICES J. Mar. Sci.*, 73(3), 849-864, doi: 10.1093/icesjms/fsv205.

Semiletov, I., Pipko, I., Gustafsson, Ö., Anderson, L. G., Sergienko, V., Pugach, S., Dudarev, O., Charkin, A., Gukov, A., Bröder, L., Andersson, A., Spivak, E., Shakhova, N., 2016. Acidification of East Siberian Arctic Shelf waters through addition of freshwater and terrestrial carbon. *Nature Geosci.*, 9, 361-365, doi: 10.1038/ngeo2695.

Qi, D., Chen, L., Chen, B., Gao, Z., Zhong, W., Feely, R. A., et al., 2016. Expansion of acidifying water in the western Arctic Ocean, *Nature Climate Change*, Accepted.

Yamamoto-Kawai, M., F. McLaughlin, and E. Carmack. 2013. Ocean acidification in the three oceans surrounding northern North America. *Journal of Geophysical Research-Oceans* 118:6274-6284, doi:10.1002/2013JC009157.

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Terrestrial Carbon Cycle

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Highlights

- Northern permafrost zone soils contain 1330-1580 billion tons organic carbon, about twice as much as currently contained in the atmosphere.
- Warming conditions promote microbial conversion of permafrost carbon into the greenhouse gases carbon dioxide and methane that are released to the atmosphere in an accelerating feedback to climate change.
- Tundra ecosystems are taking up increasingly more carbon during the growing season over the past several decades, but this has been offset by increasing carbon loss during the winter. Overall, tundra appears to be releasing net carbon to the atmosphere.

Introduction

The Arctic continues to warm at a rate that is currently twice as fast as the global average (see essay on *Surface Air Temperature*). Warming is causing normally frozen ground (permafrost) to thaw, exposing significant quantities of organic soil carbon to decomposition by soil microbes (Romanovsky et al. 2010, Romanovsky et al. 2012). This **permafrost carbon** is the remnants of plants, animals, and microbes accumulated in frozen soil over hundreds to thousands of years (Schuur et al. 2008). The northern permafrost zone holds twice as much carbon as currently in the atmosphere (Schuur et al. 2015, Hugelius et al. 2014, Tarnocai et al. 2009, Zimov et al. 2006). Release of just a fraction of this frozen carbon pool, as the greenhouse gases carbon dioxide and methane, into the atmosphere would dramatically increase the rate of future global climate warming (Schuur et al. 2013).

This report details recent advances in quantifying the amount of organic carbon stored in permafrost zone soils and recent trends (1970-2010) in the exchange of carbon between tundra ecosystems and the atmosphere. These data are the most recent comprehensive data synthesis across individual sites.

Permafrost Carbon Pools: How Much Permafrost Carbon Is Available to Release Into the Atmosphere?

Northern soils were known by scientists for decades to have relatively large amounts of carbon, accumulating in frozen and waterlogged soils (Gorham 1991). But it was only about a decade ago when attention by the science community became focused on carbon stored deeper in permafrost soils (Zimov et al. 2006), below the traditional 1-meter accounting depth. Soil carbon originates from organisms living at the surface and so organic carbon becomes increasingly scarce with depth. But soils in the northern permafrost zone have unique characteristics that can cause an accumulation of deep carbon. These characteristics include: vertical mixing due to the freeze-thaw cycle, peat accumulation as a result of waterlogged conditions, and deposits of wind and water-moved silt (**yedoma**) tens of meters thick, (Gorham 1991, Schirrmeister et al. 2002, Bockheim et al. 2007, Schuur et al. 2008).

The latest circumpolar inventory represents a synthesis of measurements from hundreds of individual study sites that now includes an order of magnitude more data for deep soils (>1m). This comprehensive inventory places the amount of organic carbon stored in the northern permafrost region at 1,330-1,580 billion tons carbon, almost twice as much carbon as currently contained in the atmosphere (**Fig. 9.1**) (Schuur et al. 2015, Hugelius et al. 2014). Of this amount, 800-1000 billion tons is perennially frozen, with the remainder contained in seasonally thawed soils. Considering just the upper 3 m of northern permafrost zone soils, the amount of carbon is equal to 50% of the quantity of soil carbon to this depth found in ecosystems everywhere else, even though this northern region represents only 15% of the global soil area (**Fig.9. 1a**) (Schuur et al., 2015).

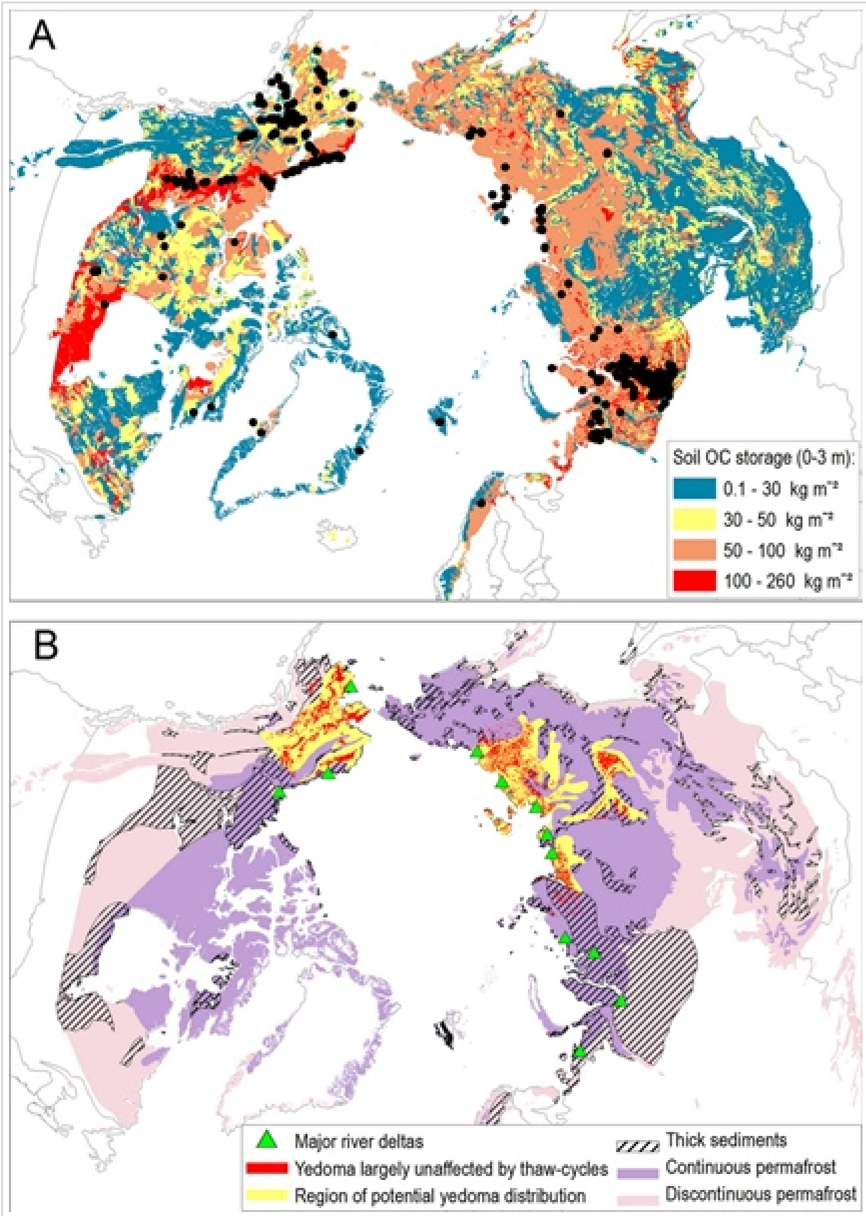

Fig. 9.1. (a) Soil organic carbon pools (to 3 m depth) for the northern circumpolar permafrost zone. Black points show locations of 0-3 m samples. (b) Deep permafrost carbon pools (>3 m) including river deltas (green triangles), and the yedoma region of Siberia and Alaska (yellow). Yedoma soils that have remained unaffected by thaw-lake cycles are shown in red. Other regions (black hashed) have thick sediments but organic carbon has not been systematically quantified. The base layer shows permafrost distribution with continuous region to the north having permafrost everywhere (>90%), and discontinuous regions further south having permafrost in some, but not all, locations (<90%) (Schuur et al. 2015).

The inventory also shows substantial permafrost carbon below 3 m depth. Deep carbon has been best quantified for the yedoma region of Siberia and Alaska, characterized by permafrost silt sediments tens of meters thick (**Fig. 9.1b**). The yedoma region covers a 1.4×10^6 km^2 area that remained ice-free during the last Ice Age (Strauss et al. 2014). The carbon inventory of this region comprises yedoma soils that were previously thawed as lakes formed and then refrozen into permafrost when lakes drained, interspersed by intact permafrost yedoma deposits that were unaffected by thaw-lake cycles (Anthony et al. 2014).

The improved inventory also highlighted additional carbon stocks that are likely to be present but are so poorly quantified that they cannot yet be added into the number reported above. There are deep soil/sediment deposits outside of the yedoma region with such limited quantification, all we can do at this point is to suggest that about 400 billion tons of additional carbon may be preserved there with unknown susceptibility to current climate change. An additional pool is organic carbon remaining in permafrost but that is now submerged on shallow Arctic sea shelves that were formerly exposed as terrestrial ecosystems during the Last Glacial Maximum ~20,000 years ago (Walter et al. 2007) (subsea area not shown in **Fig. 9.1** map). This permafrost is slowly degrading due to seawater intrusion, and it is not clear what amounts of permafrost and organic carbon still remain in the sediment versus what has already been converted to greenhouse gases.

Tundra Carbon Exchange: How Fast Can This Carbon Release Occur?

Permafrost thaw and increased microbial decomposition releases stored organic carbon from the terrestrial biosphere into the atmosphere as greenhouse gases. At the same time, plant growth sequesters atmospheric carbon, which becomes stored as new plant biomass or deposited as new soil organic matter. Climate warming can stimulate both processes, and whether Arctic ecosystems are currently a net carbon source (losses > gains) or sink (gains > losses) is an area of intense research. Ecosystem carbon balance is the relatively small difference between two large, opposing fluxes: plant carbon uptake via plant photosynthesis and growth, versus respiratory loss via metabolism by all living organisms. Across the landscape, this biological carbon cycle is then modified by physical disturbance processes such as fire, and abrupt permafrost thaw (**thermokarst**) that accelerate carbon losses while modifying rates of carbon gain. Disturbances that cause rapid carbon loss on a fraction of the landscape in any given year will accelerate change relative to that directly caused by changing climate alone.

Northern tundra ecosystems typically gain carbon (**carbon sink**) stored in plant biomass and new soil organic matter during the short summer growing season when plant photosynthesis and growth is greater than carbon respired by plants and soil back to the atmosphere. One of the most recent analyses of measurements of tundra ecosystem carbon exchange (not including high-latitude forests, fens, bogs, and mires) over the past several decades is available in Belshe et al., 2013. Belshe et al. show that the amount of carbon gain during the growing season has been increasing over time (**Fig. 9.2**). There are also more observations accumulating during this time, especially since the 1990s. Plant growth can be stimulated by warmer conditions, increased atmospheric carbon dioxide, and increased soil nutrients. Patterns of ecosystem-atmosphere carbon dioxide exchange are consistent with other observations of tundra greening and shrub expansion by satellites and aerial photography (Sturm 2001, Myers-Smith 2011) (see essay on *Tundra Greenness*).

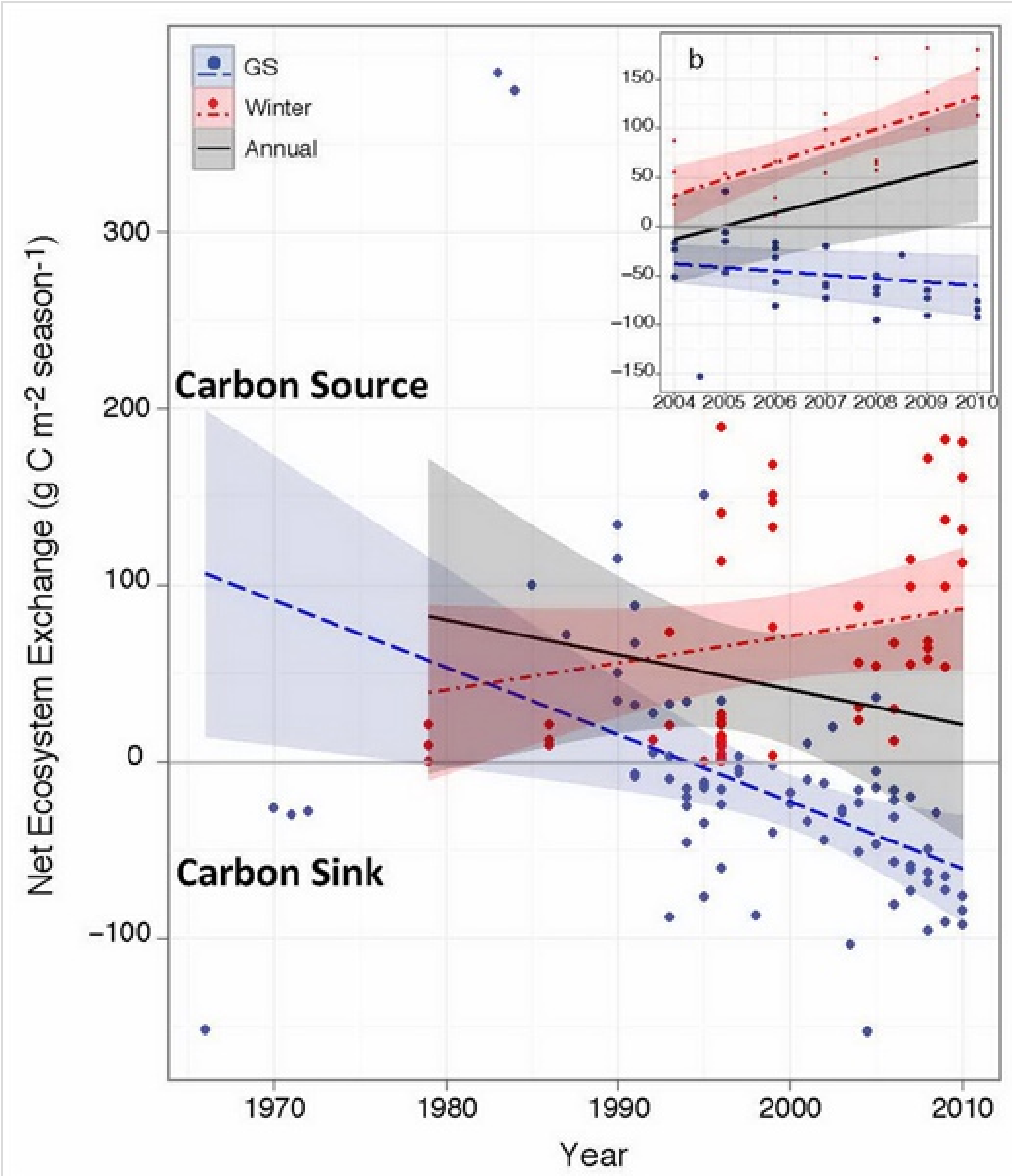

Fig. 9.2. Net exchange of carbon dioxide between tundra and the atmosphere annually (grey) and by season (GS=growing season, blue; winter, red) as a function of the year of study. Trend lines show mean and 95% confidence interval. Positive ecosystem exchange values indicate a net release of carbon to the atmosphere (Belshe et al. 2013). Inset (b) shows trend for the period 2004-2010 where winter flux datasets were in greater number.

Net carbon uptake during the growing season (June-August) is offset, at least in part, by net carbon releases during the winter (Oct-April), when carbon loss via respiration is greater than plant carbon gain (**Fig. 9.2**). Even though carbon exchange rates are generally low in the winter, the long winter season can contribute ~10-30% of annual respiration and in some cases can tip the balance between carbon sink and **carbon source**, when there is a net loss to the atmosphere. Observations of winter carbon exchange are more limited than those made during the summer but carbon release in the winter appears to be increasing over the past several decades (**Fig. 9.2**).

Summing together growing season and winter shows that tundra was acting as a carbon source on an annual basis over the decade of the 1990s into the early 2000s. Increased summer uptake may have shifted tundra towards carbon neutrality by mid-2000s (neither sink nor source; i.e. confidence intervals include zero line), but this trend is sensitive to the limited availability of winter data. As an example of this, an analysis of only the most recent time period (2004-2010) where winter flux measurements were more available, shows rapidly increasing winter carbon loss that made tundra ecosystems greater annual carbon sources to the atmosphere in the mid-late 2000s (**Fig. 9.2, inset**). The fact that the annual trend reverses when considering only the shorter, but more robust, winter dataset from 2004-2010 highlights the critical need for expanded winter carbon flux observations.

The exact same full dataset from 1970-2010 plotted versus the mean annual temperature of the study sites (rather than year of measurement) more clearly points towards tundra ecosystems acting as carbon sources during the last several decades of measurements, since the mean annual trend line with error estimate are well above the carbon neutral zero line (**Fig. 9.3**). This analysis suggests that both growing season net carbon uptake and winter net carbon loss increase with warmer mean annual climatic conditions, while the overall magnitude of carbon loss (source strength) increases as well. This pattern suggests that increased Arctic warming could further stimulate tundra to lose carbon at higher rates than observed currently, even in spite of further plant carbon uptake and greening.

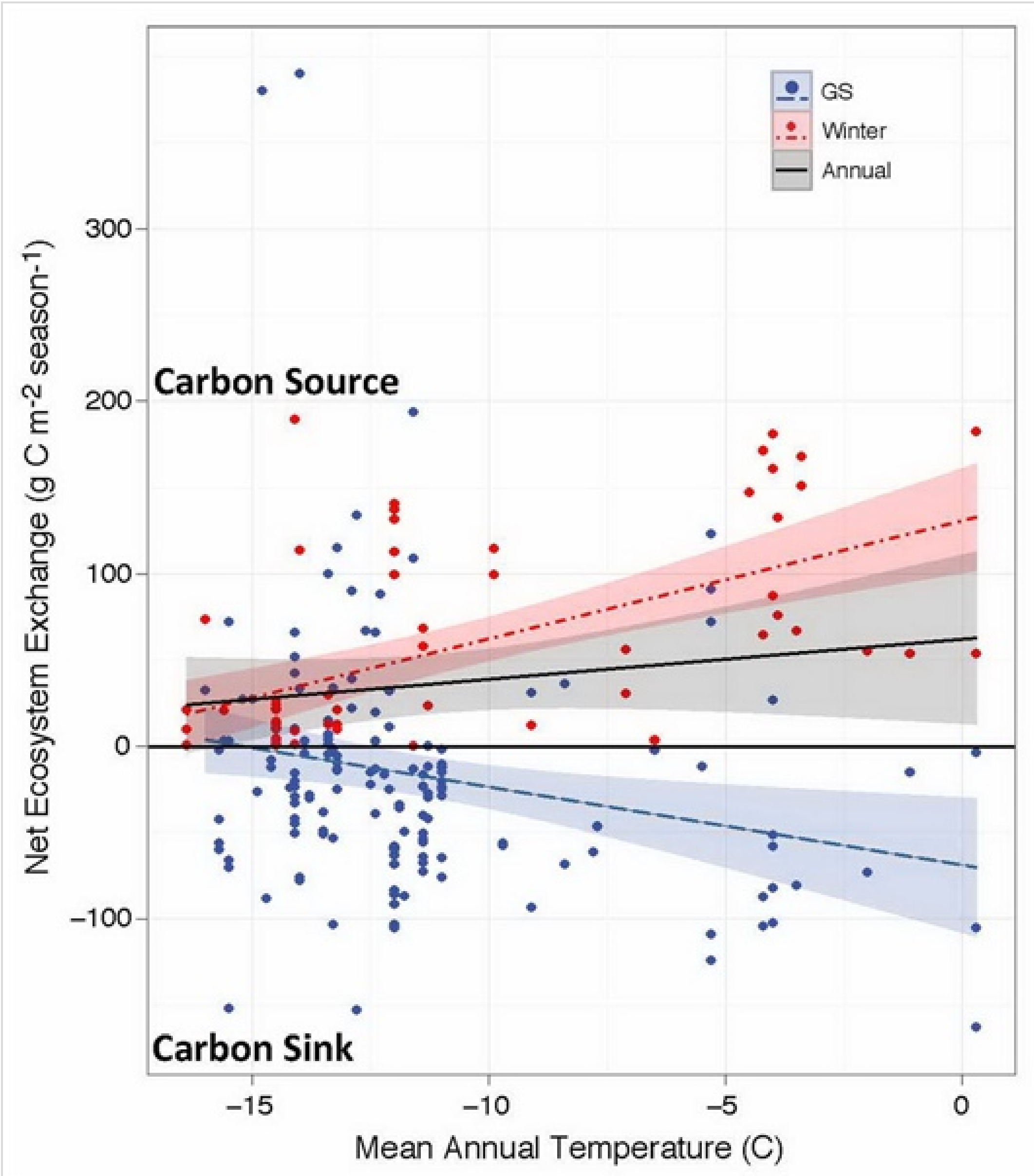

Fig. 9.3. Net exchange of carbon dioxide between tundra and the atmosphere annually (grey) and by season (GS=growing season, blue; winter, red) as a function of the mean annual temperature of the study sites using the same full 1970-2010 dataset. Trend lines show mean and 95% confidence interval. Positive ecosystem exchange values indicate a net release of carbon to the atmosphere (Belshe et al. 2013).

It is important to acknowledge that tundra ecosystem results may be influenced by where and when data have been collected, and how methodologies have changed over time. For instance, McGuire et al. (2012) conducted a related analysis of flux studies, with many studies overlapping those used in Belshe (2013) but also including wetlands. When separated into wet and dry tundra types, McGuire et al. showed mean dry tundra as an annual carbon source (though it could not be distinguished from carbon neutral), whereas mean wet tundra was an annual carbon sink.

References

Anthony, K. M. W. et al. A shift of thermokarst lakes from carbon sources to sinks during the Holocene epoch. *Nature* 511, 452, doi: 10.1038/Nature13560 (2014).

Belshe, F., E. A. G. Schuur, and B.M. Bolker. 2013. Tundra ecosystems observed to be carbon dioxide sources due to differential amplification of the carbon cycle. *Ecology Letters* doi: 10.1111/ele.12164.

Bockheim, J. 2007. Importance of cryoturbation in redistributing organic carbon in permafrost-affected soils. *Soil Science Society of America Journal* 71: 1335-1342.

Gorham, E. Northern peatlands: role in the carbon cycle and probable responses to climatic warming. *Ecological Application*s 1, 182-195 (1991).

Hugelius, G. et al. Improved estimates show large circumpolar stocks of permafrost carbon while quantifying substantial uncertainty ranges and identifying remaining data gaps. 11, 4771-4822 (2014).

Köchy M., Hiederer R., Freibauer A. 2015. Global distribution of soil organic carbon - Part 1: Masses and frequency distributions of SOC stocks for the tropics, permafrost regions, wetlands, and the world. *SOIL* 1, 351-365. doi:10.5194/soil-1-351-2015.

McGuire, A. D., T. R. Christensen, D. Hayes, A. Heroult, E. Euskirchen, J. S. Kimball, C. Koven, P. Lafleur, P. A. Miller, W. Oechel, P.Peylin, M. Williams and Y. Yi (2012) An assessment of the carbon balance of Arctic tundra: comparisons among observations, process models, and atmospheric inversions, *Biogeosciences*, 9, 3185-3204, doi:10.5194/bg-9-3185-2012.

Meyers-Smith, I. et al. 2011. Shrub expansions in tundra ecosystems: dynamics, impacts, and research priorities. *Environmental Research Letters* vol 4, no. 4.

Romanovsky, V., Smith, S. & Christiansen, H. Permafrost Thermal State in the Polar Northern Hemisphere during the International Polar Year 2007-2009: a Synthesis. *Permafrost and Periglacial Processes* 21, 106-116, doi:10.1002/ppp.689 (2010).

Romanovsky, V. E. et al. *Permafrost* [in Arctic Report Card 2012], <ftp://ftp.oar.noaa.gov/arctic/documents/ArcticReportCard_full_report2012.pdf>.

Schirrmeister, L., Siegert, C., Kunitzky, V. V., Grootes, P. M. & Erlenkeuser, H. Late Quaternary ice-rich permafrost sequences as a paleoenvironmental archive for the Laptev Sea Region in northern Siberia. *Int J Earth Sci* 91, 154-167 (2002).

Schuur, E. A. G. et al. Vulnerability of permafrost carbon to climate change: Implications for the global carbon cycle. *Bioscience* 58, 701-714 (2008).

Schuur, E. A. G. et al. Expert assessment of vulnerability of permafrost carbon to climate change. *Climatic Change* 119, 359-374, doi:10.1007/s10584-013-0730-7 (2013).

Schuur, E. A. G. et al. Climate change and the permafrost carbon feedback. *Nature* 520, 171-179, doi:10.1038/nature14338 (2015).

Strauss, J., L. Schirrmeister, G. Grosse, S. Wetterich, M. Ulrich, U. Herzschuh et al. 2013. The Deep Permafrost Carbon Pool of the Yedoma Region in Siberia and Alaska. *Geophysical Research Letters* 40, 6165-6170.

Sturm, M. C. Racine, K. Tape. Climate change: increased shrub abundance in the Arctic. *Nature* 411, 546-547 (31 May 2001) | doi:10.1038/35079180.

Tarnocai, C., Canadell, J. G., Mazhitova, G., Schuur, E. A. G. & Kuhry, P. Soil organic carbon pools in the northern circumpolar permafrost region. *Global Biogeochemical Cycles* 23 (2009).

Walter, K. M., Edwards, M. E., Grosse, G., Zimov, S. A. & Chapin, F. S. Thermokarst lakes as a source of atmospheric CH4 during the last deglaciation. *Science* 318, 633-636 (2007).

Zimov, S. A. et al. Permafrost carbon: Stock and decomposability of a globally significant carbon pool. *Geophysical Research Letters* 33 (2006).

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Shrews and Their Parasites: Small Species Indicate Big Changes

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Highlights

- As Arctic climate warms, how will terrestrial ecosystems and interconnected communities respond in the coming decades? Small mammals, such as shrews, and their parasites can serve as key indicators for anticipating the consequences of accelerating environmental perturbation and changing interactions among species.
- Some Arctic shrews have recently acquired new parasites indicating poleward shifts of sub-Arctic faunas, and demonstrating increases in both Arctic biodiversity and overall complexity within these novel species networks.
- Spatially broad and temporally deep specimen collections of small mammals and parasites are revealing the extent and nature of community reorganization as northern faunas adjust to accelerating climate change.

Introduction

When natural environments change, species may shift their distributions, adapt to novel conditions, or die out. Consequently, accelerating climate change is already recognized as leading to considerable range expansion and contraction, evolutionary changes, and extinction for hosts, parasites, and diseases through the Arctic (e.g., Kutz et al., 2013; Meltofte et al., 2013). Ultimately, these perturbations have consequences for wildlife and humans at high latitudes (e.g., Dudley et al., 2015). Complex host-parasite systems are critical proxies for understanding change in northern regions due to species interactions that determine the distribution of parasites and disease over space and time (e.g., Hoberg et al., 2012, 2013). Life histories for helminth parasites (tapeworms, flukes, roundworms) often involve circulation among mammalian hosts (where adult parasites reside) and other vertebrate and invertebrate species (where larval or developing parasites reside). Other parasites (viruses, bacteria, protozoans) circulate through vectors such as blood feeding arthropods or through direct transmission. These complex life cycles are closely tied to environmental conditions, define linkages across communities, and scale from individuals to ecosystems.

Small mammals, and especially shrews (*Sorex* spp.; **Fig. 10.1**), are diverse and relatively well documented through the Arctic (MacDonald and Cook, 2009). Shrews occupy most habitats, host a diverse and abundant parasite fauna, and function as consumers of considerable quantities of invertebrates, widely ranging from worms and crustaceans, to spiders, insects, and their larvae. Because of these connections, shrews and parasites provide direct insights into processes governing larger networks of organisms within terrestrial ecosystems (Binkienė et al., 2011; Hope et al., 2013). From shrew faunas, we now recognize how climate change through ice ages of the past two million years has facilitated repeated invasions of northern systems (e.g., Hope et al., 2015), movement between continents (Waltari et al., 2007), contact among other mammalian or avian species including exchange of parasites (Haukisalmi et al., 2010), and at times extended isolation in glacial refugia (Cook et al. 2005). We review recent advances in our understanding of the complexity of these small mammals and their parasites, emphasizing their value for interpreting generalized biodiversity responses to accelerating change and perturbation in the Arctic.

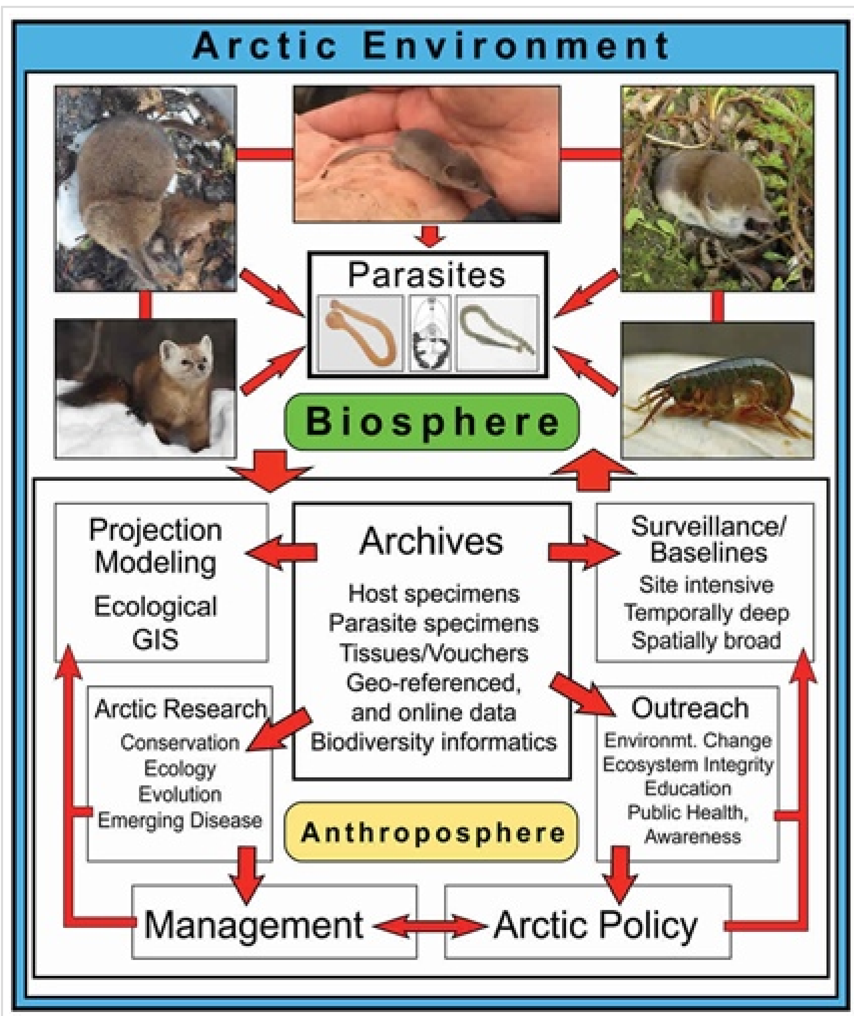

Fig. 10.1. Flow diagram for effective understanding and maintenance of integrated pathways that describe the biosphere, through biodiversity inventory, monitoring, and collection of comprehensive baselines. In this example, the biosphere is represented by shrews and other potential hosts within

associated complex parasite life cycles. This includes *Lineolepis* tapeworms, *Maritrema* flukes passing through both shrews and amphipod crustaceans where shrews are the definitive host, or *Soboliphyme* roundworms passing through marten, and with shrews in this case being an intermediate host. All of these species are in turn nested within their environment. Gradient arrows between anthroposphere (part of the environment manipulated by humans for human use) nested within the biosphere indicate a mutual interplay between these systems. Arrows are inclusive of all possible directional connections linking adjacent and remote boxes. Image credits - top left: masked shrew (A.G. Hope); top middle: Pribilof Island shrew (C. Gregory, Field Guides Inc.); top right: tundra shrew (A.G. Hope); bottom left: American marten (creative commons); bottom right: beach amphipod (buzzmarinelife.blogspot.com); middle: *Soboliphyme baturini* roundworm (A.G. Hope), *Maritrema* fluke (Karpenko and Dokuchaev, 2012), and *Lineolepis* tapeworm (V.V. Tkach).

Shrews and Parasites Reflect Generalizable Processes

Research on shrews and their parasites aims to mitigate future loss of Arctic biodiversity due to anthropogenic environmental change, habitat alteration, and novel species interactions. Persistent challenges for conservation of Arctic biodiversity include incomplete knowledge of 1) existing faunal diversity and structure, 2) ecological processes that maintain critical connections among species, and 3) processes that drive evolution (diversification and adaptation) within wild populations (Dobson et al., 2015).

Most shrew species exhibit broad continental distributions, with regional variation corresponding to major geographic barriers (e.g., Hope et al., 2012). As shrews are well represented in northern communities and are conducive to detailed specimen-based analyses, we use them to understand fundamental evolutionary and ecological processes that govern all species, from mites to muskox (Hoberg and Brooks, 2008). Genetic signatures recorded in shrew DNA can be used to understand past distributions, population sizes, and responses to environmental change. Long-term population characteristics can then be combined with climate data from the last 30 years to understand recent distributions and to project trajectories into future decades. For example, species distributional models for masked shrews (*Sorex cinereus*) associated with forests and barren ground shrews (*Sorex ugyunak*) associated with tundra indicate that the latter species is experiencing significant range contraction and fragmentation in multiple refugial areas, whereas populations of the former are expanding their distributions on northward trajectories (**Fig. 10.2**); trends that are predicted to continue into the near future.

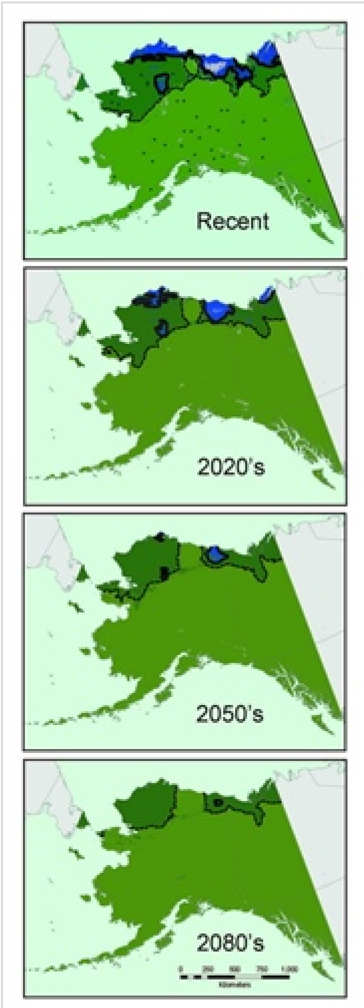

Fig. 10.2. Examples of species distributional predictions for two shrews occurring within northwestern North America. The distribution of masked shrews (*Sorex cinereus*; boreal) are shown in green and barren ground shrews (*Sorex ugyunak*; tundra) are shown in blue. Recent distributions are based on a climate normal of 30 years of record from 1981-2010. Dark green areas indicate increasing regions of overlap, contact, and interaction between these species (and their associated parasites). Dashed line indicates limits of overlap. Solid lines indicate regions where only the tundra species is predicted. Adapted from Hope et al. (2013).

Coupling genetic data (recording past responses) with climate data (governing present and future distributions) provides a powerful toolset for gauging the rate, extent, and direction of shrew responses to climate change. Our analyses indicate that Arctic adapted shrews occurring in tundra have persisted within northernmost regions through extended cold periods, often in the absence of sub-Arctic species, and with more extensive distributions than at present (**Fig. 10.3**; Hope et al., 2015). As climate has transitioned from cold to warm, our models indicate progressive contraction and fragmentation into future refugial areas of intact tundra biodiversity that are nested within historical distributions for this fauna. These refugial zones are significant in reflecting areas of continuous occupation, and a long history of ebb and flow within Arctic ecosystems. Conversely, multiple shrews species associated with boreal (forest) habitats did not endure within Arctic latitudes through cold phases of the

Pleistocene. These species have rapidly shifted ranges northward with climate warming to follow their preferred set of environmental conditions, a trend that will continue in the coming decades (Hope et al., 2013). Differential responses among forest and tundra species have resulted in range overlap and increased complexity of species interactions (Hoberg and Brooks, 2013). Support for observed shrew responses are strengthened by analysis of other mammal species, indicating concerted community-level outcomes involving northward expansion into the Arctic among forest species, and northward contraction and fragmentation of Arctic species (**Fig. 10.3**; Hoberg et al., 2012; Hope et al., 2014, 2015).

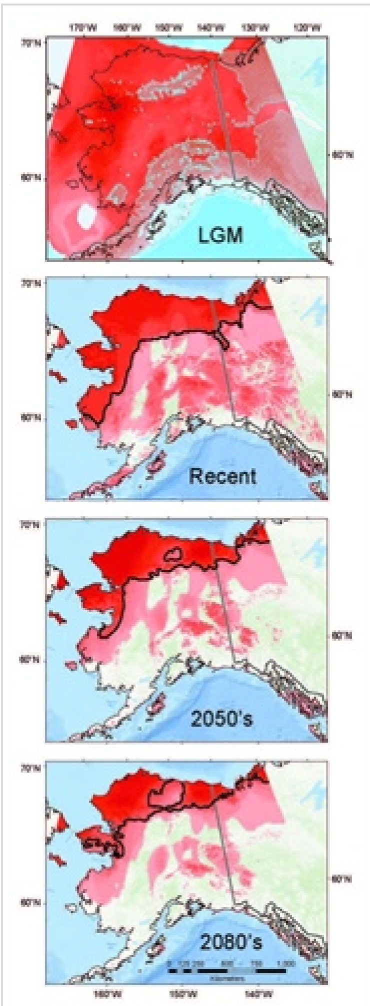

Fig. 10.3. Examples of the Arctic tundra small mammal (and parasite) community distributional predictions within northwestern North America. Combined distributions of 7 tundra small mammal species (including 1 shrew) are shown for Last Glacial Maximum (LGM; 21 thousand years ago), recent-historic, and near-future timeframes. Recent distributions are based on a climate normal of 30 years of record from 1981-2010. The color gradient reflects areas of low (light) to high (dark) diversity. Black lines provide a guideline to the extent of ≥50% of the diversity during recent and future timeframe. Blue chevrons in LGM panel indicate extent of continental ice sheets at this time. Progressive

northward contraction and fragmentation of this community has occurred in response to climate warming. Adapted from Hope et al. (2015).

Shrew parasites variably adhere to host responses through complex interactions which involve 1) geographic colonization, particularly intercontinental exchange, 2) host switching, and 3) local evolutionary adaptation to changing environments (e.g., Hoberg and Brooks, 2015). Future viability of Arctic biodiversity will be contingent on sensitivity to environmental thresholds, and relative ability to persist in the face of change, reflecting both resilience and ecological plasticity (Araujo et al., 2015). As such, although many Arctic taxa have persisted until now (through major pre-Anthropocene climate oscillations), they may be particularly vulnerable to a contemporary regime of accelerated warming not previously experienced (Serreze and Barry, 2011). Overlap among previously isolated species will create new associations which can be more readily examined within shrews and their diverse complement of parasites, compared with larger species (e.g., caribou). Hybridization between closely related species may play a critical role in diversification, adaptation, or even extinction (Hewitt, 2011). The current overlap between boreal and tundra communities in Alaska occurs in Low-Arctic tundra, where we have recently identified hybridization between shrews in these boundary zones (Hope et al., *In Prep.*). Secondary contact and host hybridization also undoubtedly contributes to mixing among parasites, intensifying these dynamics. The interplay between shrew hybridization and mixing of parasite communities result in faunal mosaics that are now a critical component being explored within a broader research agenda in the Arctic, where diverse species assemblages are undergoing complex patterns of exchange (Kelly et al., 2010; Hoberg et al., 2012).

An exemplar of shrew-parasite dynamics is being revealed within a remote island system. Islands typically have fewer species with limited ranges, lower genetic variability and reduced ability to respond to change (Pauls et al., 2013). For example, the remote island of St. Paul was isolated in the Bering Strait about 10,000 years ago and supports only a single small mammal species, the Pribilof Islands shrew (*Sorex pribilofensis*). Through collaborations with Aleuts of St. Paul and NOAA, we obtained samples of *S. pribilofensis* and confirmed that they have extremely low genetic diversity and a reduced parasite fauna dominated by a tapeworm (*Lineolepis pribilofensis*), which occurs in great abundance (sometimes hundreds per shrew).

How mammals regulate their parasites, or what conditions allow them to host diverse versus more limited parasite faunas is contingent on history and ecology, and in turn parasite diversity or intensity of infection may regulate gut microbiomes, including pathogens (Bordes and Morand, 2015). Notably, some *S. pribilofensis* specimens, in addition to being infected with a single shrew-specific tapeworm, also had heavy loads of *Maritrema* flukes. Successful infection by these helminths in shrews is remarkable considering their typical hosts are shorebirds, with parasite life cycles occurring in the intertidal zone as birds (and shrews) eat crustacean intermediate hosts such as amphipods (**Fig. 10.4**). This is an intriguing example of how island populations contribute to intercontinental transfer of parasites both between distantly related avian and mammalian hosts and through marine to terrestrial pathways via intertidal foraging, as seen elsewhere (Stewart et al., 1989).

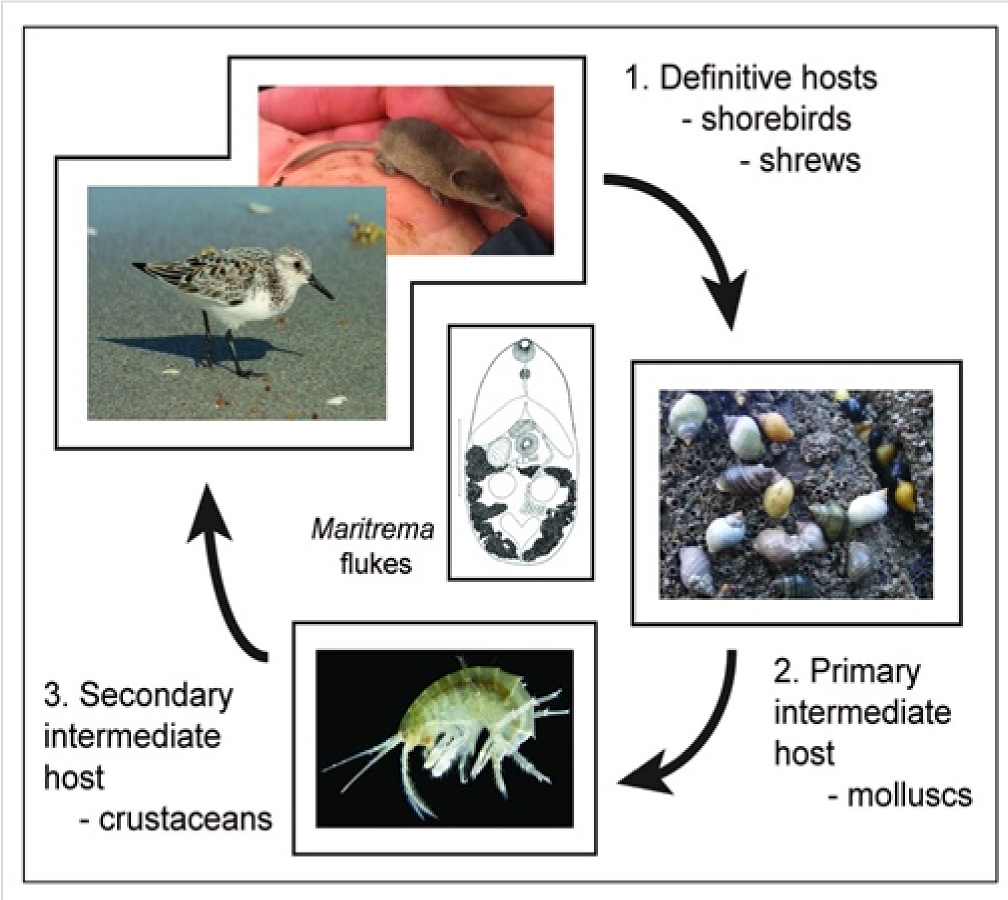

Fig. 10.4. The complex life cycle of *Maritrema* flukes in association with *Sorex pribilofensis* shrews on St. Paul Island. These parasites cycle as adults through a definitive host (usually a shorebird) which sheds eggs that hatch and infect a mollusk intermediate host, followed by a larval stage in a second intermediate crustacean host (such as an amphipod). These crustaceans are finally eaten by shorebirds (or shrews) to complete the cycle. On St. Paul, flukes found within *S. pribilofensis* inform us about shrew life history characteristics, and use of littoral zone food resources. Parasites thus provide new views of connectivity within the biosphere. Scale bar for fluke represents 0.2mm. Image credits: Sanderling shorebird (I. Sevi, Wikimedia commons); Pribilof Island shrew (C. Gregory, Field Guides Inc.); amphipod (M. Manas, Creative commons); whelks (M. Talbot, Creative commons); *Maritrema* fluke (Karpenko and Dokuchaev, 2012).

St. Paul Island is coincident with a transition zone between the Arctic and Pacific oceans, and provides another arena for complex species interactions at the intersection between marine and terrestrial ecosystems. As climate in the Bering Sea warms, this marine transition zone has shifted northward to the vicinity of St. Paul, disrupting the ecological isolation of purely Arctic communities in this area and increasing the diversity of littoral species, including amphipods (Mueter and Litzow, 2008; Weslawski et al., 2010). This shift may explain why *Maritrema* flukes now infect *S. pribilofensis*, but did not only ~50 years ago (Olsen, 1969). Such cascading effects highlight conservation implications for island species with nowhere to go, low potential to adapt, and which are now experiencing new relationships with potential pathogens; all consequences of a warming climate, ecosystems in rapid transition, and a shrinking Arctic (e.g., Hoberg et al., 2013).

Historical Archives Fuel Future Understanding

A primary issue for anticipating responses within high latitude systems continues to be a woefully incomplete knowledge of existing Arctic biodiversity (e.g., MacDonald and Cook, 2009; Hoberg et al., 2013; Cook et al., 2013). Recent compilations of shrew endoparasites (Kinsella and Tkach, 2009; Binkienė et al., 2011) highlights large gaps in our understanding of northern parasites and also documents high rates of new discoveries of parasites and viruses (Arai et al., 2008; Greiman et al., 2013; Tkach et al., 2013). However, new efforts to develop site intensive and spatially broad natural history collections are providing a rigorous foundation for multiple integrative approaches to characterize diversity and understand changing conditions (**Fig. 10.1**; McLean et al., 2015; Hoberg et al., 2015), including recognition of new shrew species and new distributional records (Cook et al., 2016). For example, genomic sequencing of historic and recent specimens is providing unparalleled views of contemporary population responses, parasite and microbiome biodiversity, and changing Arctic conditions (Cook et al., *in review*).

Current rates of Arctic environmental perturbations substantially surpass global averages (Serreze and Barry, 2011) while our ability to comprehensively understand the impacts for humans and intact wild systems continues to scratch the surface (Meltofte et al., 2013). A robust picture of diversity in the Arctic urgently requires coordinated efforts to synthesize traditional knowledge and insights from indigenous peoples, along with implementation of integrated and comprehensive field collections. An emphasis on annual targeted monitoring will provide both archival specimens and ecological information as baselines to anticipate regional and landscape responses to accelerating change.

Host-parasite systems provide a generalizable reference for understanding and documenting ecological changes (e.g., Kutz et al., 2014). Utilizing diverse, widespread, and abundant host/parasite resources from both small and large mammals, and associated communities affords critical insights about the nature and future of high latitude systems (Hoberg et al., 2012, 2013; Hope et al., 2013). As northern species have developed new associations with observable frequency through recent decades, we anticipate increasing occurrence of host switching events (and potential for emerging disease). Dynamic faunas such as shrews and their parasites embody the concept of "communities within communities" and are revealing the complexity of evolutionary and ecological processes now shaping Arctic terrestrial ecosystems.

References

Arai, S., S. N. Bennett, L. Sumibcay, J. A. Cook, J. W. Song, A. Hope, C. Parmenter, V. R. Nerurkar, T. L. Yates, and R. Yanagihara, 2008: Phylogenetically distinct hantaviruses in the masked shrew (*Sorex cinereus*) and dusky shrew (*Sorex monticolus*) in the United States. *Am. J. Trop. Med. Hyg.*, 78, 348-351.

Araujo, S. B. L., M. P. Braga, D. R. Brooks, S. Agosta, E. P. Hoberg, F. von Hathental, W. A. Boeger, 2015: Understanding host-switching by ecological fitting. *PLoS ONE*, 10, e0139225.

Binkienė, R., V. Kontrimavichus, and E. Hoberg. 2011: Overview of the cestode fauna of European shrews of the genus Sorex with comments on the fauna in *Neomys* and *Crocidura* and an exploration of historical processes in post-glacial Europe. *Helminthologia*, 48, 207-228.

Bordes, F., and S. Morand, 2015: Impacts of parasite diversity on wild vertebrates: limited knowledge but important perspectives, Ch. 5. In: (Eds. S. Morand, B. R. Krasnov, and D. T. L. Littlewood) *Parasite Diversity and Diversification: Evolutionary Ecology Meets Phylogenetics*, Cambridge University Press, Cambridge, United Kingdom.

Cook, J. A., E. P. Hoberg, A. Koehler, H. Henttonen, L. Wickström, V. Haukisalmi, K. Galbreath, F. Chernyavski, N. Dokuchaev, A. Lahzuhtkin, S. O. MacDonald, A. Hope, E. Waltari, A. Runck, A. Veitch, R. Popko, E. Jenkins, S. Kutz, and R. Eckerlin, 2005: Beringia: intercontinental exchange and diversification of high latitude mammals and their parasites during the Pliocene and Quaternary. *Mamm. Study*, 30, S33-S44.

Cook, J., C. Brochman, S. L. Talbot, V. Fedorov, E. Taylor, R. Väinölä, E. P. Hoberg, M. Kholodova, M. Magnusun, 2013: Genetics, pp. 514-539 In: (Ed. H. Meltofte) *Arctic Biodiversity Assessment- Status and Trends in Arctic Biodiversity*. Conservation of Arctic Flora and Fauna, Arctic Council, Akureyi, Iceland.

Cook, J. A., B. S. McLean, D. J. Jackson, J. P. Colella, S. E. Greiman, V. V. Tkach, T. S. Jung, and J. L. Dunnum, 2016: First record of the Holarctic least shrew (*Sorex minutissimus*) and associated helminths from Canada: new light on northern Pleistocene refugia. *Can. J. Zool.*, 94, 367-372.

Dobson, A., P. K. Molnár, and S. Kutz, 2015: Climate change and Arctic parasites. *Trend. Parasitol.*, 31, 181-188.

Dudley, J. P., E. P. Hoberg, E. J. Jenkins, A. J. Parkinson, 2015: Climate change in the North American Arctic: A One Health perspective. *EcoHealth*, 12, 713-725.

Greiman, S. E., V. V. Tkach, and J. A. Cook, 2013: Description and molecular differentiation of a new *Staphylocystoides* (Cyclophyllidea: Hymenolepididae) from the dusky shrew *Sorex monticolus* in Southeast Alaska. *J. Parasitol.* 99, 1045-1049.

Haukisalmi, V., L. M. Hardman, P. Foronda, C. Feliu, J. Laakkonen, J. Niemimaa, J. T. Lehtonen, and H. Henttonen, 2010: Systematic relationships of hymenolepidid cestodes of rodents and shrews inferred from sequences of 28S ribosomal RNA. *Zoologica Scripta*, 39, 631-641.

Hewitt, G. M., 2011: Quaternary phylogeography: the roots of hybrid zones. *Genetica*, 139, 617-638.

Hoberg, E. P., and D. R. Brooks, 2008: A macroevolutionary mosaic: Episodic host-switching, geographic colonization, and diversification in complex host-parasite systems. *J. Biogeog.*, 35, 1533-1550.

Hoberg, E. P., and D. R. Brooks, 2013: Episodic processes, invasion, and faunal mosaics in evolutionary and ecological time, pp. 199-213 In: (Ed. K. Rohde) *The Balance of Nature and Human Impact*. Cambridge University Press.

Hoberg, E. P., and D. R. Brooks, 2015: Evolution in action: Climate change, biodiversity dynamics and emerging infectious disease. Theme Issue- Climate Change and Vector-Borne Diseases. *Phil. Trans. R. Soc. B*, 2015: 370, 20130553.

Hoberg, E. P., K. E. Galbreath, J. A. Cook, S. J. Kutz, and L. Polley, 2012: Northern host-parasite assemblages: History and biogeography on the borderlands of episodic climate and environmental transition, pp. 1-97 In: (Eds. D. Rollinson and S.I. Hays) *Advances in Parasitology*, 79, Elsevier.

Hoberg, E. P., S. J. Kutz, J. A. Cook, K. Galaktionov, V. Haukisalmi, H. Henttonen, S. Laaksonen, A. Makarikov, and D. J. Marcogliese, 2013: Parasites in terrestrial, freshwater and marine systems, pp. 476-505 In: (Ed. H. Meltofte) *Arctic Biodiversity Assessment- Status and Trends in Arctic Biodiversity*. Conservation of Arctic Flora and Fauna, Arctic Council, Akureyi, Iceland.

Hoberg, E. P., S. J. Agosta, W. A. Boeger, and D. R. Brooks, 2015: An Integrated parasitology: Revealing the elephant through tradition and invention. *Trend. Parasitol.*, 31, 128-133.

Hope, A. G., K. A. Speer, J. R. Demboski, S. L. Talbot, and J. A. Cook, 2012. A climate for speciation: Rapid spatial diversification within the *Sorex cinereus* complex of shrews. *Mol. Phylogenet. Evol.*, 64, 671-684.

Hope, A. G., E. Waltari, D. C. Payer, J. A. Cook, and S. L. Talbot, 2013: Future distribution of tundra refugia in northern Alaska. *Nature Clim. Change*, 3, 931-938.

Hope, A. G., S. W. Y. Ho, J. L. Malaney, J. A. Cook, and S. L. Talbot, 2014a: Accounting for rate variation among lineages in comparative demographic analyses. *Evolution*, 68, 2689-2700.

Hope, A. G., E. Waltari, J. L. Malaney, D. C. Payer, J. A. Cook, and S. L. Talbot, 2015. Arctic biodiversity: increasing richness accompanies shrinking refugia for a cold-associated tundra fauna. *Ecosphere*, 6, 1-67.

Karpenko, C. V., and N. E. Dokuchaev, 2012. The first record of the trematode *Maritrema afanassjewi* Belopolskaja, 1952 in shrews. Vestnik Dalnevostochnogo Otdeleniya Rossiyskoj Akademii Nauk, 163, 103-106. (in Russian)

Kelly, B., D. A. Tallmon, and A. Whiteley, 2010: The Arctic melting pot. *Nature*, 468, 891.

Kinsella, J. M., and V. V. Tkach, 2009: Checklist of helminth parasites of Soricomorpha (= Insectivora) of North America north of Mexico. *Zootaxa*, 1969, 36-58.

Kutz, S. J., S. Checkley, G. G. Verocai, M. Dumond, E. P. Hoberg, R. Peacock, J. Wu, K. Orsel, K. Seegers, A. Warren, and A. Abrams, 2013: Invasion, establishment, and range expansion of two parasitic nematodes in the Canadian Arctic. *Global Change Biol.*, 19, 3254-3262.

Kutz, S., E. P. Hoberg, P. K. Molnár, A. Dobson, and G. Verocai, 2014: A walk on the tundra: Host-parasite interactions in an extreme environment. *Internat. J. Parasitol.- Parasites Wildlife*, 3, 198-208.

MacDonald, S. O. and J. A. Cook, 2009: Recent Mammals of Alaska. University of Alaska Press, Fairbanks. 387pp.

McLean, B. S., K. C. Bell, J. L. Dunnum, B. Abrahamson, J. P. Colella, E. R. Deardorff, J. A. Weber, A. K. Jones, F. Salazar-Miralles, and J. A. Cook, 2015: Natural history collections-based research: progress, promise, and best practices. *J. Mamm.*, p. 178.

Meltofte, H., T. Barry, D. Berteaux, H. Bultmann, J. S. Christiansen, J. A. Cook, A. Dahlberg, F. J. A. Daniëls, D. Ehrich, J. Fjeldså, F. Fridriksson, B. Ganter, A. J. Gaston, L. J. Gillespie, L. Grenoble, E. P. Hoberg, I. D. Hodkinson, H. P. Huntington, R. A. Ims, A. B. Josefson, S. J. Kutz, S. L. Kuzmin, K. L. Laidre, D. R. Lassuy, P. N. Lewis, C. Lovejoy, C. Michel, V. Mokievsky, T. Mustonen, D. C. Payer, M. Poulin, D. G. Reid, J. D. Reist, D. F. Tessler, and F. J. Wrona, 2013: Arctic Biodiversity Assessment. Conservation of Arctic Fauna and Flora Committee, Copenhagen.

Mueter, F. J. and M. A. Litzow, 2008: Sea ice retreat alters the biogeography of the Bering Sea continental shelf. *Ecol. App.*, 18, 309-320.

Olsen, O. W., 1969: *Hymenolepis pribilofensis* n. sp. (Cestoda: Hymenolepididae) from the Pribilof shrew (*Sorex pribilofensis* Merriam) from the Pribilof Islands, Alaska. *Can. J. Zool.*, 47, 449-454.

Pauls, S. U., C. Nowak, M. Bálint, and M. Pfenninger, 2013: The impact of global climate change on genetic diversity within populations and species. *Mol. Ecol.*, 22, 925-946.

Serreze, M. C. and R. G. Barry, 2011. Processes and impacts of Arctic amplification: A research synthesis. *Glob. Planet. Change*, 77, 85-96.

Stewart, D. T., T. B. Herman, and T. Teferi, 1989. Littoral feeding in a high-density insular population of *Sorex cinereus*. *Can. J. Zool.*, 67, 2074-2077.

Tkach, V. V., A. A. Makarikov, and J. M. Kinsella, 2013: Morphological and molecular differentiation of *Staphylocystis clydesengeri* n. sp. (Cestoda, Hymenolepididae) from the vagrant shrew, *Sorex vagrans* (Soricimorpha, Soricidae) in North America. *Zootaxa*, 3691, 389-400.

Waltari, E., E. P. Hoberg, E. P. Lessa, and J. A. Cook, 2007: Eastward Ho: phylogeographical perspectives on colonization of hosts and parasites across the Beringian nexus. *J. Biogeog.*, 34, 561-574.

Weslawski J. M., J. Wiktor Jr., and L. Kotwicki, 2010: Increase in biodiversity in the arctic rocky littoral, Sorkappland, Svalbard, after 20 years of climate warming. *Marine Biodiv.*, 40, 123-130.

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Arctic Change - So What?: Linkages and Impacts

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The Arctic is an integral part of the larger Earth system where multiple interactions unite its natural and human components. As is amply demonstrated in each annual installment of the Arctic Report Card, the domain is collectively experiencing rapid and amplified signatures of global climate change. At the same time, the Arctic system's response to this broader forcing has, itself, become a central research topic, given its potential role as a critical throttle on future planetary dynamics (NRC 2013, 2014). Changes are already impacting life systems, cultures and economic prosperity and continued change is expected to bear major implications far outside the region (ACIA 2005, AMAP 2012, IPCC 2013, Cohen et al. 2014). Ongoing assessments of how the system is wired-together and how sensitive its environment is to change suggest that there are important interconnections and possible feedbacks but these remain highly uncertain (Francis et al. 2009a; Hinzman et al. 2013). We have entered an era when environmental management, traditionally local in scope, must confront regional, whole biome, and pan-Arctic challenges but also requires policy development that crosses scales and boundaries from villages to international partnerships.

Because of these rapid, if not unprecedented, rates of change, the Arctic is today home to a suite of literally trillion dollar impacts—both positive and negative. Examples are as varied as global trade and the opening of new trans-Arctic shipping routes, increased or impeded access to land and ocean-based resources, changing ecosystems and fisheries, upheaval in subsistence resources, damage to infrastructure as on fragile coastlines, Arctic sovereignty and national security concerns, climate change adaptation and mitigation. While issues may appear to exist in isolation, they are emerging very much within the context of an evolving, integrated Arctic system defined by interactions among its major natural and social sub-systems. By 'system' we mean the constellation of major and subsidiary components residing within the Arctic atmosphere, ocean, land that comprise and, through interacting processes, define the physical, biological, chemical and energetic state of the Arctic that vary over both space and time.

In the U.S., a recent raft of strategic planning documents point to the Arctic as a key arena of national concern; from science, policy and private sector perspectives. Recent administrations have recognized the importance of the region, as evidenced by several high-level planning documents and executive orders (President Bush 2009; President Obama 2014, 2015, IARPC 2015). US Arctic Research Commission Goals and Objectives Reports, which help to prioritize agency and interagency investments in Arctic research, have in recent years reported on several systems-level challenges (e.g. the apparent paradox of increasing precipitation with simultaneous drying across the Arctic landmass). Accelerating U.S. interest in the Arctic is reflected in its Chairmanship of the Arctic Council (2015-17) and its four themes, which are all arguably systems-level issues: Arctic Ocean safety, security, and stewardship; building Arctic community resilience; recognizing and responding to climate change; raising Arctic awareness as a fundamental part of the Earth's natural systems and economy.

A systems-level understanding is clearly necessary to articulate the role of the region in the broader Earth-system (the fundamental science question) and in the global economy (a key societal question). Such understanding is an important precursor to identify future trajectories of the Arctic, evaluating its resiliency and long-term sustainability. Systems-level understanding is also required to assess both the short-term and legacy impacts of specific human decisions, by analyzing the feedbacks they invoke on natural, physical, social, and economic sub-systems and the still newer decisions that may possibly be contemplated (or actually executed) in response. Relevant examples include management of wildlife populations through hunting regulations and the consequent impacts on plant biomass (Russell and Gunn, 2012 and Joly and Klein, 2016), with plausible additional impacts on permafrost or even microbial dynamics that feedback to emission of radiatively important gases like methane. While curiosity-based research will be fundamental to progress, a science agenda additionally co-generated through a partnership among scientists, policymakers and stakeholders will help ensure that we develop the tools and understanding that can address evolving societal concerns. The capacity to analyze the "decision-impact-next decision" space has yet to be developed and would represent an excellent partnership of scientific knowledge providers and consumers.

One ideal example of how a systems perspective is absolutely essential to addressing both a scientific and societal challenge is the ever-present issue of community relocation as a result of rapidly eroding Arctic coastlines (Forbes et al. 2011). Although the dilemma of threatened coastal communities goes well beyond the Arctic, the interconnected processes and their common sources traceable back to climate warming are distinct from those in more temperate regions. Beach erosion in the continental U.S. is generally related to changing sediment sources, suspension and deposition, and complicated by rising seas, increased frequency of storminess, and associated storm surge. The integrity of Arctic shorelines is similarly compromised, but additionally by the loss of protection afforded by disappearing sea ice, which normally tempers both wave action and the effect of storms, and by coastal permafrost degradation. The Earth systems challenge thus requires an understanding of meteorology, sea ice dynamics, oceanography, and permafrost dynamics.

A relocating community is also critically defined by cascading human actions. Although in many cases, immediate action is of paramount concern (Alaska IAWG 2009), delays are more the norm (GAO 2009) and arguably arise from insufficient knowledge of indigenous and modern social systems function. Until very recently, villagers lived a more mobile lifestyle, dictated by environmental change or subsistence pressures, and resided in semi-permanent homes easily abandoned and reconstructed elsewhere. With the benefits of modern infrastructure like schools, clinics, washeterias, electrical, water delivery and sewer systems, relocating today has become a much more imposing and expensive challenge. Communities site themselves according to some geographical advantage, access to protection or food resources, and associated cultural traditions. And while alternative sites may share the same intrinsic advantages, they may suffer similar, though unforeseen vulnerabilities from future erosion. Successful repositioning of communities also requires knowledge of formal governance systems (e.g. when securing federal support for shoreline protection and assistance in relocation), which requires astute leadership at both federal and local levels.

From the systems perspective we can see these as linked cultural-environmental issues, in the arena of social science, indigenous knowledge, economics, law, community management and federal bureaucracy. The socio-environmental challenges must be integrated with civil, mechanical and coastal engineering analyses and project design, all of which depend upon the geological, hydrological, geophysical, meteorological, climatological, oceanographic, and sea ice sciences. Attempting to resolve community problems without considering the full human-environment system is likely to lead to failure, with consequences that could last for generations.

Despite its organization around reports of individual Arctic sub-domains, the Arctic Report Card itself contains many intriguing examples of change reported in its time series that could be interpreted in the context of system processes and feedbacks. The 2015 Report Card, for example, showed that melt dynamics of the Greenland Ice Sheet cannot be well explained without reference to atmospheric pressure anomalies and thus ocean-atmosphere connections associated with the North Atlantic Oscillation (NAO) (Tedesco et al. 2015). The ongoing loss of sea ice, while progressive and of great concern to scientists and policymakers, is at the same time creating positive and negative changes in ocean biology, including net primary production, but a full appreciation requires knowledge on how these changes play out with respect to position along shelves, freshwater stratification, nutrient upwelling, changes in cloudiness, etc. (Frey et al. 2015). These questions are further complicated by considerations of increased river flows, glacial melting, and ocean acidification. Broad-scale tundra "greening", recently transformed to "browning", requires an understanding of permafrost-water relations, plant-herbivore interactions, the trapping of blowing snow and an interpretation of how these macro-system changes relate to microcosm experimental responses to heating (Epstein et al. 2015). Duffy et al. (2005) demonstrated a link between atmospheric-ocean variability and the severity of the Alaskan fire season. Consequent impacts on caribou and migratory waterfowl are likely to play a role in future approaches to wildlife management.

The research community has over the last decade progressed substantially with the observational underpinnings that could be used to decipher major processes of the Arctic system (witness the Report Card, Hinzman et al. 2013, Jeffries et al. 2013). In addition, there are many new studies into the fundamental mechanics of system sub-components, for example, improved understanding of sea ice dynamics (Kwok and Untersteiner, 2011), functioning of marine (Kedra et al. 2015) and terrestrial ecosystems (Raynolds et al. 2014), permafrost hydrology (Liljedahl et al. 2016) and climate (Richter-Menge and Mathis, 2016). Furthermore, there are calls from within the Arctic agency funding community noting the marked absence of systems-level proposals to date (Swanberg and Holmes, 2013). And, perhaps most importantly, there is a demand from policymakers requesting more comprehensive visions of Arctic change (Balton, 2014). We have yet to put these productive "forces" into action to motivate a more complete understanding of system connectivity and interdependencies and where the Arctic will progress in the future, with or without human interventions. We see that the time is ripe to catalyze systems-focused studies.

What might be some of the approaches to systems-level analysis? The Arctic research, education, and policy communities have been developing new perspectives aimed at systems-level questions, including ARCSS-funded synthesis activities like the Big Sky workshop (Overpeck et al. 2005), planning within the ARCSS Committee and a consensus dialogue begun in 2005 by >100 Arctic natural and social scientists, technologists, educators, policy and outreach experts (ARCUS 2007). NSF-ARCSS SIMS and OPERA were programs designed to support Arctic systems-oriented thinking and synthesis. By their very nature, SEARCH and ISAC (Murray et al. 2010, ARCUS 2014) are motivated by Arctic systems research themes and integrated modeling and observational approaches. Synthesis studies like Francis et al. (2009) and Overpeck et al. (2005) have taken the form of explorations, environmental "Gedanken thought experiments" exploring potential linkages and feedbacks, while Hinzman et al. (2013) explored the implications of Arctic system change drawn from literature-based integration of observations and model results. Posing research questions beyond the reach of individual researchers (NRC, 2014) also can reveal higher-level emergent properties of systems. Recognizing that systems exist at multiple scales, how such processes accumulate (or not) from subsidiary scales to full system behaviors is a critical research challenge (e.g. how plot-level terrestrial biogeochemistry is consistent with pan-Arctic carbon fluxes, USARC 2010). Attribution studies can be used decipher causal chains associated with natural "experiments", for example, Eurasian River discharge anomalies linked to an extreme springtime melt event (e.g., Rawlins, 2009).

Recognizing that the Arctic is a coupled water-energy-biogeochemical system embodies the notion of "currencies" that define budgets, fluxes, feedbacks and hence facilitate the study of system behaviors. For example, the water cycle is linked to energy cycling through the energy required for phase changes, to the carbon (C) cycle through controls on CO_2 and CH_4 in ecosystems, and to the nitrogen (N) cycle through controls on its transport and availability for plants and microbes. In turn, water, energy, and N are linked to C by limiting tissue stoichiometry and plant growth. These cycles are all changing concurrently across the Arctic, with change in one component reverberating strongly into others (e.g. sea ice loss interacting with Arctic vegetation [Bhatt et al. 2010] or the polar jet stream [Francis et al. 2009b]). Finally, high resolution, full-system Arctic system models (Roberts et al. 2010) have been argued for. In the context of policymaker engagement, however, simpler models (of intermediate complexity) (Eby et al. 2013) could also be useful, with the advantage of being far less computationally intensive to set-up and run and hence enable co-design of scenarios with policymakers with rapid turnaround on "what-if" questions.

In addition, we see frameworks that support systems research as an important Arctic science infrastructure investment. We speak here of a collaboratory concept (Bos et al. 2007) to forward systems perspectives, methods, and tools for analysis of multi-scalar, multi-dimensional, trans-disciplinary issues in the Arctic. An Arctic Systems Collaboratory by its very nature would constitute a networked community resource, organized as an interactive virtual or actual environment to execute basic research as well as to co-design a science agenda with the Arctic stakeholder community.

In conclusion, meeting the major scientific challenges associated with understanding and forecasting the future state of the Arctic system will require balanced advances in (1) the scientific state-of-the-art and (2) the mechanisms to overcome the traditional disciplinary boundaries, which—in our view—have slowed earlier progress. One interesting example of the existence of such boundaries is this very report. We recognize the Arctic Report Card as an important synthesis of observations that can be analyzed on their own terms (inductive studies) but also critical for model parameterization, validation, establishing initial conditions and boundary forcings (deductive/modeling approaches). We see, however, an important missed opportunity to convey the essence of Arctic change, namely Arctic *system* change. For this reason, we recommend that the Report Card move purposefully into the domain of systems-level issues. With a small augmentation of its content, it could begin making strides toward breaking down the disciplinary boundaries reflected in its current chapter orientation. One structure might be to feature, along the lines of the examples presented earlier, short pieces on land-atmosphere, ocean-atmosphere, and land-ocean connections. Given the synoptic character of the Report, it has become an essential part of the science-to-policy arsenal, and we make our recommendation in the spirit of increasing its public policy impact.

References

ARCUS, 2007: *The Arctic Synthesis Collaboratory: An Integrated Community-based Research Network*. Arctic Collaboratory Planning Committee & ARCSS Committee, eTown Meeting, 5 Dec 2007, http://archive.arcus.org/arcss/etm/december_07/ArcticCollaboratory_ETown_Dec07FINAL.ppt.

ARCUS, 2014: *Reports and Science Plans, Study of Environmental Arctic Change*. Available at http://www.arcus.org/search-program/resources/reports-and-science-plans.

ACIA, 2005: *Arctic Climate Impact Assessment*. Cambridge University Press, New York, NY 1042 pp.

AMAP, 2012: *Arctic Climate Issues 2011: Changes in Arctic Snow, Water, Ice and Permafrost*. SWIPA 2011 Overview Report. Arctic Monitoring and Assessment Programme (AMAP). Oslo, xi, 97 pp.

Balton, D. A. 2014: *Statement of Ambassador David A. Balton, Deputy Assistant Secretary for Oceans and Fisheries, Bureau of Oceans and International Environmental and Scientific Affairs, U.S. Department of State*. Before the Committee on Transportation and Infrastructure, Subcommittee on Coast Guard and Maritime Transportation. U.S. House of Representatives 23 Jul 2014.

Bhatt, U. S., D. A. Walker, M. K. Raynolds, J. C. Comiso, H. E. Epstein, G. J. Jia, R. Gens, J. E. Pinzon, C. J. Tucker, C. E. Tweedie, and P. J. Webber, 2010: Circumpolar Arctic tundra vegetation change is linked to sea-ice decline. *Earth Interactions*. 14, 1-20, doi:10.1175/2010EI315.1.

Bos, N., A. Zimmerman, J. Olson, J. Yew, J. Yerkie, E. Dahl, and G. Olson, 2007: Shared Databases to Communities of Practice: A Taxonomy of Collaboratories. *Journal of Computer-Mediated Communication*, 12, 652-672.

Cohen, J., J. A. Screen, J. C. Furtado, M. Barlow, D. Whittleston, D. Coumou, J. Francis, K. Dethloff, D. Entekhabi, J. Overland, and J. Jones, 2014: Recent Arctic amplification and extreme mid-latitude weather. *Nature Geoscience*, 7, 627-637.

Duffy, P. A., J. E. Walsh, J. M. Graham, D. H. Mann, and T. S. Rupp, 2005: Impacts of large-scale atmospheric-ocean variability on Alaskan fire season severity. *Ecological Applications*. 15, 1317-1330.

Eby, M., A. J. Weaver, K. Alexander, K. Zickfeld, et al., 2013: Historical and idealized climate model experiments: An intercomparison of Earth system models of intermediate complexity. *Climate of the Past*, 9 (3), 1111-1140, doi: 10.5194/cp-9-1111-2013.

Epstein et al., 2015: Tundra Greenness. *Arctic Report Card*, http://www.arctic.noaa.gov/Report-Card.

Forbes, D. L. (editor). 2011: *State of the Arctic Coast 2010 - Scientific Review and Outlook*. International Arctic Science Committee, Land-Ocean Interactions in the Coastal Zone, Arctic Monitoring and Assessment Programme, International Permafrost Association. Helmholtz-Zentrum, Geesthacht, Germany, 178 p. http://arcticcoasts.org.

Francis, J. A., D. M. White, J. J. Cassano, W. J. Gutowski Jr., L. D. Hinzman, M. M. Holland, M. A. Steele, and C. J. Vörösmarty, 2009a: An arctic hydrologic system in transition: Feedbacks and impacts on terrestrial, marine, and human life. *J Geophys. Res.*, 114, G04019, doi:10.1029/2008JG000902.

Francis, J. A., W. Chan, D. Leathers, J. R. Miller, and D. E. Veron, 2009b: Winter Northern Hemisphere weather patterns remember summer Arctic sea ice extent. *Geophys. Res. Lett.*, 36, L07503, doi:10.1029/2009GL037274.

Frey et al., 2015: Arctic Ocean Primary Productivity. *Arctic Report Card*. http://www.arctic.noaa.gov/Report-Card.

GAO. 2009: Relocation of Alaska Native Villages: Limited Progress Has Been Made on Relocating Villages Threatened by Flooding and Erosion. *Report to Congressional Requesters*. GAO-09-551, Government Accounting Office, Washington, DC.

Hinzman, L. D., C. J. Deal, A. D. McGuire, S. H. Mernild, I. V. Polyakov, and J. E. Walsh, 2013: Trajectory of the Arctic as an Integrated System. *Ecological Applications*, 23, 1837-1868.

IAWG, 2009: *Alaska's IAWG 2009 Final Report: Recommendations to the Governor's Sub-Cabinet on Climate Change*, 162p, climatechange.alaska.gov/docs/iaw_finalrpt_12mar09.pdf.

IPCC, 2013: *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*. T. F. Stocker et al., Eds. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 1535 pp.

Jeffries, M. O., J. E. Overland, and D. K. Perovich, 2013: The Arctic shifts to a new normal. *Phys. Today*, 66(10), 35 (2013); doi: 10.1063/PT.3.2147, http://dx.doi.org/10.1063/PT.3.2147.

Joly, K. and D. R. Klein, 2016: Complexity of Caribou Population Dynamics in a Changing Climate. *U.S. National Park Service*. https://www.nps.gov/articles/aps-v10-i1-c7.htm.

Kedra, M., C. Moritz, E. S. Choy, C. David, R. Degen, S. Duerksen, I. Ellingsen, B. Gorska, J. M. Grebmeier, D. Kirievskaya, D. van Oevelen, K. Piwosz, A. Samuelsen, J. M. Wesławski, 2015: Status and trends in the structure of Arctic benthic food webs. *Polar Research*, 34, 23775, http://dx.doi.org/10.3402/polar.v34.23775.

Kwok, R. and N. Untersteiner, 2011: The thinning of Arctic sea ice, *Phys. Today*, 64(4), 36-41.

Liljedahl, A. K., J. Boike, R. P. Daanen, A. N. Fedorov, G. V. Frost, G. Grosse, L. D. Hinzman, Y. Iijma, J. C. Jorgenson, N. Matveyeva, M. Necsoiu, M. K. Raynolds, V. E. Romanovsky, J. Schulla, K. D. Tape, D. A. Walker, C. J. Wilson, H. Yabuki, D. Zona., 2016: Pan-Arctic ice-wedge degradation in warming permafrost and its influence on tundra hydrology. *Nature Geoscience*, 9,312-318, doi:10.1038/ngeo2674.

Murray, M. S., L. Anderson, G. Cherkashov, C. Cuyler, B. Forbes, J. C. Gascard, C. Haas, P. Schlosser, G. Shaver, K. Shimada, M. Tjernström, J. Walsh, J. Wandell, and Z. Zhao, 2010: International Study of Arctic Change: Science Plan. *ISAC International Program Office*, Stockholm. 58 pp.

National Research Council, 2013: *Abrupt Impacts of Climate Change: Anticipating Surprises. Polar Research Board, Committee on Understanding and Monitoring Abrupt Climate Change and its Impacts, Board on Atmospheric Sciences and Climate*, J. W. C. White, R. B. Alley, D. Archer, A. D. Barnosky, J. Foley, R. Fu, M. Holland, S. Lozier, J. Schmitt, L. C. Smith, G. Sugihara, D. Thompson, and A. Weaver. The National Academies Press, Washington, DC, 208 pp.

National Research Council, 2014: *The Arctic in the Anthropocene: Emerging Research Questions. Polar Research Board, Committee on Emerging Research Questions in the Arctic*. H. Huntington, S. Pfirman, C. Ashjian, L. Bourgeau-Chavez, J. A. Francis, S. Haakanson, R. Hawley, T. Hepa, D. Hik, L. Hinzman, A. Lynch, A. M. Macrander, G. Miller, K. Moran, E. Mosley-Thompson, S. Mukasa, and T. Weingartner. The National Academies Press, Washington, DC, 220 pp.

Overpeck, J. T., M. Sturm, J. A. Francis, D. K. J. Perovich, M. C. Serreze, R. Benner, E. C. Carmack, F. S. Chapin III, S. C. Gerlach, L. C. Hamilton, M. Holland, H. P. Huntington, J. R. Key, A. H. Lloyd, G. M. MacDonald, J. A. McFadden, D. Noone, T. D. Prowse, P. Schlosser, and C. Vörösmarty, 2005: Arctic system on trajectory to new, seasonally ice-free state. *Eos Transactions*, 86 (34), 309.

President George W. Bush, 2009: *US Arctic Policy*. National Security Presidential Directive/NSPD-66, Homeland Security Presidential Directive/HSPD-25.

President Barack Obama, 2014: *Implementation Plan for The National Strategy for the Arctic Region*.

President Barack Obama, 2015: *Executive Order - Enhancing Coordination of National Efforts in the Arctic*. http://www.whitehouse.gov/the-press-office/2015/01/21/executive-order-enhancing-coordination-national-efforts-arctic.

Rawlins, M. et al., 2009: Tracing freshwater anomalies through the air-land-ocean system: A case study from the Mackenzie River basin and the Beaufort Gyre. *Atmos.-Ocean*, 47, 79-97.

Raynolds, M. K., D. A. Walker, K. J. Ambrosius, J. Brown, K. R. Everett, M. Kanevskiy, G. P. Kofinas, V. E. Romanovsky, Y. Shur, and P. J. Webber, 2014: Cumulative geoecological effects of 62 years of infrastructure and climate change in ice-rich permafrost landscapes, Prudhoe Bay Oilfield, Alaska. *Global Change Biology*, 20,1211-1224. DOI: 10.1111/gcb.12500.

Richter-Menge, J. and J. Mathis, Eds., 2016: The Arctic [in "State of the Climate in 2015"]. *Bull. Amer. Meteor. Soc.*, 97(8), S131-S154.

Roberts, A., L. D. Hinzman, J. E. Walsh, M. Holland, J. Cassano, R. Döscher, H. Mitsudera, and A. Sumi, 2010: A Science Plan for Regional Arctic System Modeling: A report by the Arctic Research Community for the National Science Foundation Office of Polar Programs. International Arctic Research Center Technical Paper 10-0001. *International Arctic Research Center*, University of Alaska Fairbanks, 47 pp.

Russell, D. and A. Gunn, 2012: Caribou and Reindeer (Rangifer). *Arctic Report Card: Update for 2012*. ftp://ftp.oar.noaa.gov/arctic/documents/ArcticReportCard_full_report2012.pdf.

Swanberg, N. and R. M. Holmes, 2013: Is it ARCSS? *Witness the Arctic*, 17(3), 4-6.

Tedesco, M. et al., 2015: Greenland Ice Sheet. Arctic Report Card, http://www.arctic.noaa.gov/Report-Card.

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Faster Glaciers and the Search for Faster Science

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During a 1979 research cruise in the Bering Sea, Conrad Oozeva, a Native hunter from St. Lawrence Island, shared dozens of Yupik words for sea ice (**Fig. 12.1**). I recently looked at my notebook from that period and realized that some of those terms—such as *tagneghneq* (thick, dark, weathered ice)—refer to types of sea ice that are rare or non-existent today. That some of those Yupik terms—probably in use for thousands of years—would become obsolete in just a few decades attests to the rapid pace of change in the Arctic and to the impacts on indigenous peoples (Berman 2004; Oozeva et al. 2004; Ford and Pearce 2010).

Fig. 12.1. Brendan Kelly (author) and Conrad Oozeva aboard the USCGC *Polar Sea* in 1979.

For Arctic researchers, communicating the impacts of our discoveries has taken on an unprecedented urgency in the face of environmental change that—in many instances—is outpacing our ability to understand and explain the changes we are witnessing. Accustomed to advancing our scientific disciplines at what is often called a 'glacial' pace, we recognize that glaciers are not so slow anymore (Arendt et al. 2002; Howat et al. 2008; Schild et al. 2016). Before long, we may need to redefine 'glacial' to mean something that is rapidly diminishing or employ a different adjective. Glacial melts are accelerating, and—to serve society—we need to accelerate our science. Of the many steps in creating new scientific knowledge, some are more amenable to acceleration than others. The final step—communicating new knowledge—certainly needs to be sped up.

Scientists working in the Arctic and elsewhere increasingly recognize the importance of becoming more timely and effective in conveying what we know. And, what we know often is most valuable to policy makers and affected communities when we synthesize across disciplines and succinctly communicate the policy-relevant points. For example, researchers are combining observations and models in oceanography, atmospheric sciences, sea ice physics, and glaciology to understand the impacts of diminishing sea ice on mid-latitude weather (Rahmstorf and Coumou 2011; Francis and Vavrus 2012, 2015; Swain et al. 2016) and the melting of glaciers and ice sheets (Rahmstorf et al. 2015). A series of cross-disciplinary workshops have accelerated the pace of synthesis (https://usclivar.org/meetings/2017-arctic-midlatitude-workshop) by providing important and timely fora for translation among scientists across disciplines. Yet, we still need further translation-and condensation-to convey the state of the science to policy makers (https://www.youtube.com/watch?v=5eDTzV6a9F4). Accelerating our science means speeding up syntheses and improving our ability to effectively convey what we know.

In the United States, the science community has increasingly recognized the value of making scientific results broadly accessible since 1997 when the National Science Foundation adopted the description of broader impacts as one of the criteria for funding. Subsequently, organizations such as COMPASS (http://www.compassonline.org/OurHistory), the Leopold Leadership Program (http://leopoldleadership.stanford.edu/about/mission), the Union of Concerned Scientists (http://www.ucsusa.org/action/science_network/science-network-workshop-series.html#.V-7LWZMrJsM), and the American Association for the Advancement of Science (https://www.aaas.org/pes/communicating-science-workshops) developed effective trainings that help scientists communicate more effectively with diverse audiences. Science agencies and organizations, including NOAA (http://www.arctic.noaa.gov/Report-Card/), NASA (https://science.nasa.gov/science-news), and the American Geophysical Union (https://sharingscience.agu.org/) now provide fora for disseminating digests of new science.

The evolution of Study of Environmental Arctic Change (SEARCH) is a good example of how Arctic scientists increasingly are accelerating their science and communication. SEARCH was founded to advance understanding of the Arctic system and its trajectory through synthesis and modeling. From a scientific perspective, synthesis is an important step in an iterative process. It "is the critical last step before the design of the next experiment" (Patrick Webber quoted in Overpeck 2003). At its inception, SEARCH recognized the need to include a focus on synthesis (Morison et al. 1998). That early vision of SEARCH described observing and process studies at length but devoted less than 10% of the text to synthesis. That disparate treatment reflected, perhaps, the program's nascency. Five years later, NSF sponsored the Arctic System Science Synthesis Retreat in Big Sky, Montana that began with a careful contrast between synthesis and analysis and resulted in the projection that the Arctic is headed for a "super interglacial state" (Overpeck et al. 2005).

Inspired by the Big Sky synthesis, SEARCH's early focus on the interaction of ocean and atmosphere (Morison et al. 1998; Wang and Key 2003; Liu et al. 2005; Wang and Key 2005a,b) expanded to facilitating syntheses considering freshwater (Serreze et al. 2006; Holland et al. 2007; White et al. 2007; Francis et al. 2009; Rawlins et al. 2010), terrestrial ecosystems (Hinzman et al. 2005), marine ecosystems (Huntington et al. 2014), connections between the Arctic and the broader earth system (Serreze et al. 2000; Chapin et al. 2005; Huntington et al. 2007a; Francis and Vavrus 2012), and human dimensions of change in the Arctic (Huntington et al. 2007b; 2015). Other recent examples of Arctic syntheses efforts include reports on the evolving science linking Arctic sea ice loss to changes in mid-latitude weather (National Academy of Sciences 2014; Jung et al. 2015) and marine ecosystems (Bellard et al. 2012; Carmack et al. 2012; Post et al. 2013).

SEARCH and many in the Arctic research community have become increasingly convinced, however, that additional important syntheses combining scientific research and Indigenous Knowledge are needed. From a scientific perspective, synthesis may be important for designing the next experiment, but policy makers, local communities, and others look to syntheses to answer specific questions. Syntheses framed in policy-relevant forms (e.g., how much and how soon will sea level rise? Or, how soon will coastal erosion undermine our village?) are more valuable to these user communities than are framings focused on advancing the state of knowledge. And, the policy questions are becoming increasingly urgent.

SEARCH now brings together scientists as well as stakeholders and agencies to synthesize knowledge from many disciplines—syntheses intended to simultaneously increase the body of knowledge and address stakeholder questions. Recognizing it is insufficient to say, "Science isn't done until the journal article is published," SEARCH also believes that science isn't done until stakeholders use the science.

Translating science into forms usable by stakeholders' calls for translating technical information into language accessible to diverse audiences. SEARCH recognizes that many interested audiences are sophisticated but not facile with technical jargon. To improve communication with all stakeholders, we are developing "knowledge pyramids." Each knowledge pyramid assembles the state of the science concerning a societally important Arctic question in multiple formats ranging from one-page, jargon-free summaries at the apex of the pyramid (https://www.arcus.org/search-program/products) to original research publications at the base (**Fig. 12.2**). Thus, when asked about the state of the science concerning—for example—melting ice sheets and their impact on sea level rise, we would point a geologist to primary literature at the base of the pyramid; a scientist from another discipline to a review article (mid pyramid); a science journalist to a more condensed synthesis (e.g., an *Arctic Report Card* essay); and a Congressional staffer to a briefing paper in the apex. Naturally, the level at which someone enters the pyramid is not fixed and will vary with their specific background and interests. Especially important in this regard is the potential for the one-page summaries to be useful not only for policy makers but also for efficient communication among scientists of different disciplines. We believe that giving specialists windows in to each other's science will facilitate the multi-disciplinary collaborations necessary for a fuller understanding of environmental change in the Arctic. We would argue further that translating our research in to common language deepens our own understanding of our results and their broader implications.

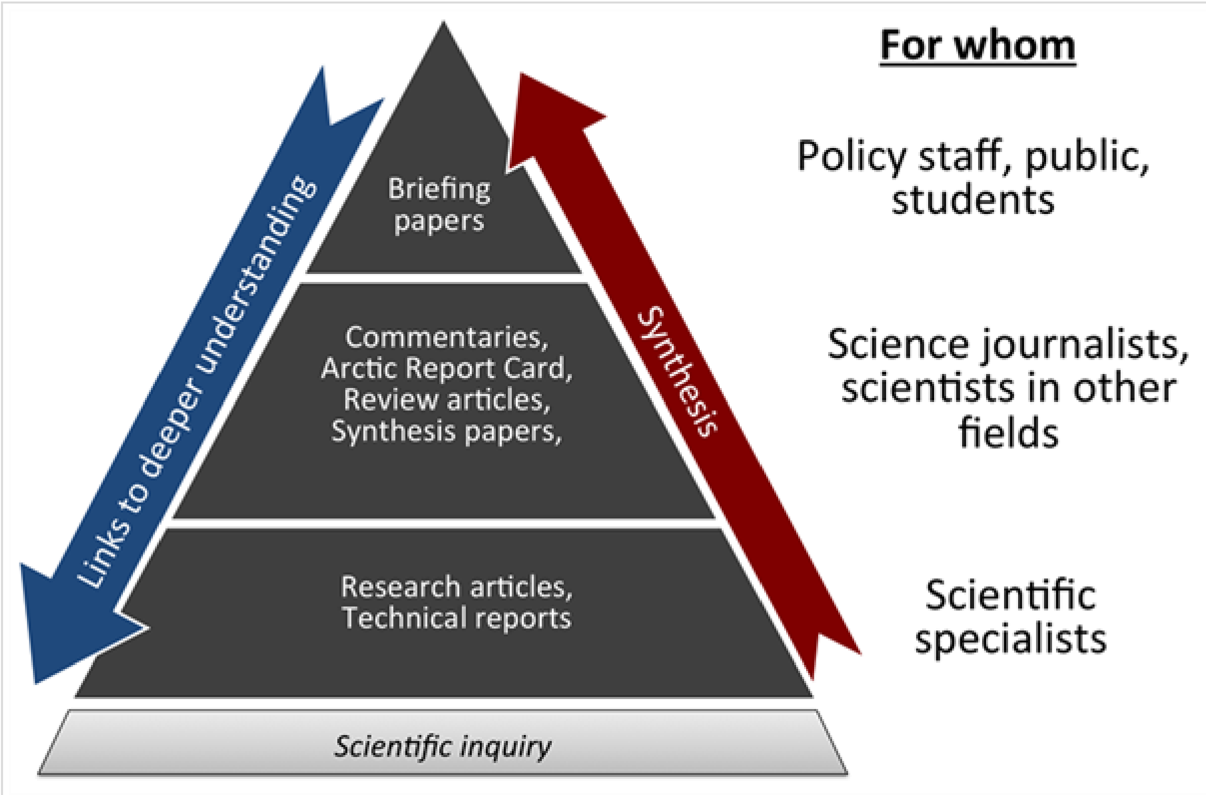

Fig. 12.2. SEARCH's concept of knowledge pyramids.

Similarly, we fully appreciate and honor the valuable information found in the differences between scientific and indigenous perceptions of the Arctic. When Conrad used numerous Yupik words to describe sea ice, which I would have referred to using a single term, he drew my attention to differences in ice characteristics that I had overlooked. Conrad's knowledge extends well beyond nouns; he is deeply knowledgeable about the exchange of energy and materials in the bio geophysical environment, as I learned when he and I co-taught a course in ecology. Our pupils were high school students in a large wall tent on St. Lawrence Island. Early on, they expressed anxiety over choosing between Conrad's indigenous knowledge and my scientific knowledge. Conrad, however, disabused them of the notion that they faced a dilemma. He eloquently described how learning from his elders, his own observations, and the many scientists he had guided around the island expanded his understanding of his environment. You can apply all those types of information in your own understanding, he explained. Thus, he encouraged our students to synthesize.

The communities of St. Lawrence Island, like communities across the Arctic, are facing extremely rapid changes, some of which may make obsolete certain terms in their language. Such cultural losses may challenge those communities, but following Conrad's advice to synthesize—drawing on information from various sources—will likely enhance their resilience. The scientific community can also benefit from Conrad's advice to think across disciplines and his example of translating his knowledge for diverse audiences.

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References

Arendt, A. A., Echelmeyer, K. A., Harrison, W. D., Lingle, C. S., and Valentine, V. B. 2002. Rapid wastage of Alaska glaciers and their contribution to rising sea level. *Science* 297:382-386.

Bellard, C., C. Bertelsmeier, P. Leadley, W. Thuiller, and F. Courchamp, 2012. Impacts of climate change on the future of biodiversity. *Ecology Letters* 15: 365-377 doi: 10.1111/j.1461-0248.2011.01736.x.

Berman, M. 2004. Adaptation and Sustainability in a Small Arctic Community: Results of an Agent-Based Simulation Model. *Arctic* 57:401-414.

Carmack, E. C., Whiteman, G., Homer-Dixon, H., and McLaughlin, F. A. 2012. Detecting and coping with potentially disruptive shocks and flips in complex-adaptive Arctic marine systems: A resilience approach to place and people, *Ambio*, 41, 56-65, DOI 10.1007/s13280-011-0225-6.

Chapin, F. S., M. Sturm, M. C. Serreze, J. P. McFadden, J. R. Key, A. H. Lloyd, A. D. McGuire, T. S. Rupp, A. H. Lynch, J. P. Schimel, J. Beringer, H. E. Epstein, L. D. Hinzman, G. Jia, C. -L. Ping, K. Tape, W. L. Chapman, E. Euskirchen, C. D. C. Thompson, J. M. Welker, and D. A. Walker. 2005. Role of land surface changes in Arctic summer warming, *Science*, 310, doi:10.1126/science.1117368, October 28.

Ford, J. D., and T. Pearce. 2010. What we know, do not know, and need to know about climate change vulnerability in the western Canadian Arctic: a systematic literature review. *Environ. Res. Lett.* 5 014008.

Francis, J. A., D. M. White, J. J. Cassano, W. J. Gutowski Jr., L. D. Hinzman, M. M. Holland, M. A. Steele, and C. J. Vörösmarty. 2009. An arctic hydrologic system in transition: Feedbacks and impacts on terrestrial, marine, and human life, *J. Geophys. Res.*, 114, G04019, doi:10.1029/2008JG000902.

Francis, J. A., and S. J. Vavrus, 2012. Evidence linking Arctic amplification to extreme weather in midlatitudes. *Geophys. Res. Lett.*, 39, LO6801, doi:10.1029/2012GL051000.

Francis, J. A., and S. J. Vavrus, 2015. Evidence for a wavier jet stream in response to rapid Arctic warming. *Environ. Res. Lett.*, 10, doi:10.1088/1748-9326/10/1/014005.

Hinzman, L. et al. 2005. Evidence and Implications of Recent Climate Change in Northern Alaska and Other Arctic Regions. *Climatic Change* 72(3):251-298.

Holland, M. M., J. Finnis, A. P. Barrett, and M. C. Serreze. 2007. Projected changes in Arctic Ocean freshwater budgets, *J. Geophys. Res.*, 112, G04S55, doi:10.1029/2006JG000354.

Howat, I. M., I. Joughin, M. Fahnstock, B. E. Smith, and T. Scambos. 2008. Synchronous retreat and acceleration of southeast Greenland outlet glaciers 2000-06: ice dynamics and coupling to climate. *Journal of Glaciology*, 54:646-660.

Huntington, H. P., M. Boyle, G. E. Flowers, J. W. Weatherly, L. C. Hamilton, L. Hinzman, C. Gerlach, R. Zulueta, C. Nicolson, and J. Overpeck. 2007a. The influence of human activity in the Arctic on climate and climate impacts. *Climatic Change* 82:77-92.

Huntington, H. P., L. C. Hamilton, C. Nicolson, R. Brunner, A. Lynch, A. E. J. Ogilvie, and A. Voinov. 2007b.Toward understanding the human dimensions of the rapidly changing arctic system: insights and approaches from five HARC projects. *Regional Environmental Change* 7(4):173-186.

Huntington, H. P., E. Carmack, P. Wassmann, F. Wiese, E. Leu, and R. Gradinger. 2015. A new perspective on changing Arctic marine ecosystems: panarchy adaptive cycles in pan-Arctic spatial and temporal scales. In: S. Arico, ed. *Ocean sustainability in the 21st century*. Cambridge, UK: Cambridge University Press. p. 109-126.

Jung, T., Doblas-Reyes, F., Goessling, H., Guemas, V., Bitz, C., Buontempo, C., Caballero, R., Jakobson, E., Jungclaus, J., Karcher, M., Koenigk, T., Matei, D., Overland, J., Spengler, T., Yang, S. 2015. Polar-lower latitude linkages and their role in weather and climate prediction. In: International Workshop on Polar-lower Latitude Linkages in Weather and Climate Prediction, 10-12 December 2014, Barcelona, Spain. *Bull. Am. Meteorol. Soc.* http://dx.doi.org/10.1175/BAMS-D-15-00121.1.

Morison, J., K. Aagaard, and M. Steele. 1998. Report on Study of the Arctic Change Workshop held November 10-12, 1997. University of Washington, Seattle, Washington. Report No. 8.

National Academy of Sciences. 2014. Linkages between Arctic Warming and Midlatitude Weather Patterns. The National Academies Press, Washington, DC (2014).

Oozeva, C., C. Noongwook, G. Noongwook, C. Alowa, and I. Krupnik. 2004 Watching ice and weather our way / Akulki, Tapghahmii, Mangtaaquli, Sunqaanga, Igor Krupnik. Sikumengllu Esghapalleghput. Edited by Igor Krupnik, Henry Huntington, Christopher Koonooka, and George Noongwook. Washington, DC: Arctic Studies Center, Smithsonian Institution. 208pp.

Overpeck, J. T. 2003. ARCSS Synthesis Retreat. Presentation at Arctic System Science (ARCSS) Synthesis Retreat. Big Sky, Montana. August 2003.

Overpeck, J. T., M. Sturm, J. A. Francis, D. K. Perovich, M. C. Serreze, R. Benner, E. C. Carmack, F. S. Chapin III, S. C. Gerlach, L. C. Hamilton, L. D. Hinzman, M. Holland, H. P. Huntington, J. R. Key, A. H. Lloyd, G. M. MacDonald, J. McFadden, D. Noon, T. D. Prowse, P. Schlosser, and C. Vörösmarty. 2005. Arctic System on Trajectory to New State, *EOS*, 86(34), 309-314.

Post, E., U. S. Bhatt, C. M. Bitz, J. F. Brodie, T. L. Fulton, M. Hebblewhite, J. Kerby, S. J. Kutz, I. Stirling, and D. A. Walker, 2013. Ecological Consequences of Sea-Ice Decline. *Science* 341:519-524.

Rahmstorf, S., J. E. Box, G. Feulner, M. E. Mann, A. Robinson, S. Rutherford, and E. J. Schaffernicht, 2015. Exceptional twentieth-century slowdown in Atlantic Ocean overturning circulation, *Nat. Clim. Change*, doi:10.1038/NCLIMATE2554.

Rahmstorf, S., and D. Coumou, 2011. Increase of extreme events in a warming world. *Proc. Natl. Acad. Sci. USA*, 108, doi: 10.1073/pnas.1101766108.

Rawlins, M. A. et al. 2010. Analysis of the Arctic System for Freshwater Cycle Intensification: Observations and Expectations. *J. Climate*. 23:5715-5737.

Schild, K. M., R. L. Hawley, and B. F. Morriss. 2016. Subglacial hydrology at Rink Isbræ, West Greenland inferred from sediment plume appearance. *Annals of Glaciology*, pp. 1-10. doi: 10.1017/aog.2016.1.

Serreze, M. C., J. E. Walsh, F. S. Chapin, III, T. Osterkamp, M. Dyurgerov, V. Romanovsky, W. C. Oechel, J. Morison, T. Zhang, and R. G. Barry. 2000. Observational evidence of recent change in the northern high latitude environment. *Climate Change* 46:159-207.

Serreze, M. C. et al. 2006. The large-scale freshwater cycle of the Arctic. *J. Geophys. Res.*, 111, C11010, doi:10.1029/2005JC003424, 2006.

Swain, D. L., D. E. Horton, D. Singh, and N. S. Diffenbaugh, 2016. Trends in atmospheric patterns conducive to seasonal precipitation and temperature extremes in California. *Sci. Adv.* 2, doi:10.1126/sciadv.1501344.

Wang, X. and J. Key. 2003. Recent trends in Arctic surface, cloud, and radiation properties from space, *Science*, 299(5613):1725-1728.

Wang, X. and J. Key. 2005a. Arctic surface, cloud, and radiation properties based on the AVHRR Polar Pathfinder data set. Part I: Spatial and temporal characteristics. *J. Climate* 18(14):2558-2574.

Wang, X. and J. Key. 2005b. Arctic surface, cloud, and radiation properties based on the AVHRR Polar Pathfinder data set. Part II: Recent trends, *J. Climate*. 18(14):2575-2593.

White, D., et al. (2007). The arctic freshwater system: Changes and impacts, *J. Geophys. Res.*, 112, G04S54, doi:10.1029/ 2006JG000353.

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