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The changing extent of the glaciers along the western Ross Sea, Antarctica

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ABSTRACT

We examine the change in terminus positions of glaciers flowing into the western Ross Sea, Antarctica, between 71°S and 78°S as a proxy for changes in snowfall and/or summer air temperature. This coastline's major glaciers terminate in ice tongues, which are particularly sensitive to changes in flow rate and calving. Using historic maps and satellite imagery spanning 60 yr (A.D. 1955–2015), the terminus positions, ice speed, calving rates, and ice front advance and retreat rates for 34 glaciers are documented. Additionally, changes in regional ice flow speed from 2008 to 2014 are examined. No significant spatial or temporal patterns of terminus position, flow speed, or calving emerged, implying that the conditions associated with ice tongue stability are unchanged. A weak trend of advance may be present in the northernmost part of the coast, consistent with estimates of increased snow accumulation and glacier mass balance in that region. The stability of these glaciers over the past half century contrasts sharply with the rapidly shrinking glaciers of the Antarctic Peninsula and suggests that no significant climate change, as manifest in glacier change, has reached this region of Antarctica.

INTRODUCTION

From A.D. 1950 to 1998, the atmosphere over the Antarctic Peninsula was warming faster than the global average (Oliva et al., 2017), and it has been cooling slightly since (Oliva et al., 2017; Turner et al., 2016). In response, the glacial landscape there has been shrinking: rapidly in the years prior to 1998, and more slowly since (Oliva et al., 2017). Ice shelves, formed from glaciers that flow out onto the ocean, have been retreating, and in some cases catastrophically disintegrating, in recent decades (Scambos, et al., 2000, 2014; Cook et al., 2005). Glacier thicknesses have declined, with magnitudes near the terminus averaging –0.28 m yr^{-1} since the mid-1960s and as much as –6 m yr^{-1} since the 1990s (Kunz et al., 2012). The rates of glacier mass loss since 2001 have totaled ~–0.7 m yr^{-1} water equivalent (Scambos et al., 2014).

In contrast, air temperatures over East Antarctica during the past half century either have shown little change (Turner et al., 2005) or are perhaps warming slowly (Steig et al., 2009; Screen and Simmonds, 2012). Seasonal sea ice in the Ross Sea has been expanding (Parkinson and Cavalieri, 2012; Simmonds, 2015), and the alpine glaciers of the McMurdo Dry Valleys have been in equilibrium (Fountain et al., 2016). The glacier termini in the Ross Island area over the period 1973–2001 have shown little change except for parts of the McMurdo Ice Shelf, which retreated in places as much as 5 km (Ferrigno et al., 2010). Glaciers along the west coast of the Ross Sea have shown a cyclic behavior of advance and retreat since 1956 with no significant trend with time (Frezzotti, 1993, 1997). Farther north, near Cape Adare, little historic change in the position of glacier termini has been observed (Frezzotti and Polizzi, 2002; Miles et al., 2013).

In light of the extensive shrinkage of glaciers on the Antarctic Peninsula over the past two decades, here we reexamine and update the variation of glacier extent along the west coast of the Ross Sea. Our analysis also includes ice speeds and calving rates, which control terminus position.

The eastern portion of the East Antarctic Ice Sheet flows through the Transantarctic Mountains to the western Ross Sea coast. This coast spans >6° of latitude (>700 km) from McMurdo Sound, 77.85°S, to Cape Adare, 71.17°S (Fig. 1). It is mostly glacier covered with local alpine glaciers and major outlet glaciers from the ice sheet, with many of the larger glaciers forming floating ice tongues. We identified 34 glaciers with clearly defined termini in the ocean, making them suitable for our analysis.

METHODS

The locations of glacier termini were determined from digital scans of paper maps and from satellite imagery. The maps were U.S. Geological Survey 1:250,000-scale maps based on aerial imagery acquired from 1955 to 1970 (U.S. Geological Survey, 1965, 1968, 1967, 1969, 1970, 1972). The satellite imagery came from a variety of platforms: Landsat, ASTER (Advanced Spaceborne Thermal Emission and Reflection Radiometer), QuickBird, and WorldView (Table DR1 in the GSA Data Repository[1]). Because of slight differences in georectification, all images

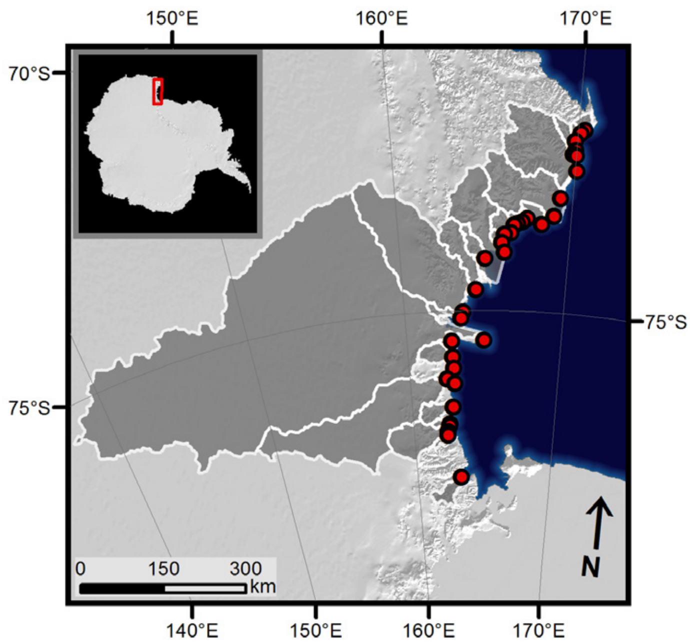

Figure 1. Glaciers of western Ross Sea (Antarctica) from Cape Adare south to Ferrar Glacier near Ross Island. Gray shading represents catchment area of each glacier within this study that feeds an ice tongue (red circle). Flow divides, which define catchment boundaries, are estimated from digital elevation model with ~200 m cell size (Liu et al., 2001). Coastal outline was obtained from Mouginot et al. (2017).

[1]GSA Data Repository item 2017310, Figures DR1–DR7 and Tables DR1–DR5, is available online at http://www.geosociety.org/datarepository/2017/ or on request from editing@geosociety.org.

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were georeferenced to a 2008–2012 mosaic (Polar Geospatial Center, 2014) because of its high spatial resolution and comprehensive coverage. Control points were ice-free bedrock features yielding root mean square error in horizontal location of 15 m or less. A polar stereographic projection was used (referenced to 71°S, projected from the WGS84 ellipsoid).

Terminus positions were digitized manually. To define a single value for the terminus location, five parallel lines in the flow direction, fixed in space and equidistant from each other and away from the side margin, were drawn over the glacier termini (Fig. DR1 in the Data Repository). These lines served as guides to measure the terminus position on sequential images. The intersection of the lines with the termini yielded five points for each glacier. The distance between the same points in sequential images was measured and averaged to determine the terminus change. Speed of the floating ice tongue was estimated by averaging the displacement of four to five features on the glacier surface, typically large crevasses, over the known time interval between images. Glacier width, defined along a line perpendicular to the glacier flow near the grounding line, is the average of widths measured on each image. Location of the grounding line was defined by Bindschadler et al. (2011).

The uncertainty of terminus position and ice speed was calculated from the square root of the sum of squares of uncertainty associated with georeferencing and image resolution for the satellite images; for the maps, it also included width of the line defining the glacier boundary. The displacement uncertainty between two images is the square root of the sum of uncertainties for each image or map.

Estimates of ice flux of each glacier used surface speed at the grounding line and glacier width. Surface speed was taken from an ice velocity data set (Rignot et al., 2011) using an image-derived grounding line location (Bindschadler et al., 2011). Ice thicknesses were not available for most of the glaciers. However, there is an approximate linear correlation between thickness and width for the subset of glaciers with thickness measurements. Because our interest is relative differences in discharge rather than the discharge magnitude, we estimated a proxy flux using the two variables best known, speed and width (units of $m^2\ yr^{-1}$). Changes in the ice speed across the region and at the grounding line were determined by differencing the ice velocity data of Rignot et al. (2011), based on 2008 imagery, from those of Fahnestock et al. (2016) using austral "summer" imagery spanning October 2013 to March 2014.

RESULTS

A total of 49 maps and images from 1955 to 2015 were digitized (Table DR1). Thirty-four (34) glaciers with marine outlets were identified with catchment areas ranging from 39 km^2 to 162,303 km^2 (Table DR2). All glaciers terminated in ice tongues extending beyond the coastline, except the two southernmost (the Ferrar Glacier and an unnamed glacier). The largest is David Glacier, which feeds the Drygalski Ice Tongue, covering almost twice the area of all other glaciers combined. Image quality was generally good, clearly defining the glacier termini and major crevasses. The early imagery was of lower spatial resolution, and for some periods the imagery was either not available or too clouded, preventing terminus definition or crevasse tracking.

Between 1955 and 2015, 19 of the 34 glaciers advanced, 13 retreated, and two were missing data (Table DR3). To compare the advance and retreat data, given different time periods, the data were normalized for time and expressed as a rate of change in m yr^{-1}. The average rate of change of all glaciers was $+12 \pm 88$ m yr^{-1} (± standard deviation). The largest observed advance was the David Glacier–Drygalski Ice Tongue, (ID 23 in Table DR2; 162,303 km^2), $+392 \pm 164$ m yr^{-1}. The greatest observed decrease was at Nordenskjold Glacier (ID 28; 23,141 km^2), -191 ± 164 m yr^{-1}. David Glacier–Drygalski Ice Tongue also exhibited the fastest speed during any period, 866 m yr^{-1} (2008–2012), and had the fastest average speed overall, 711 m yr^{-1}. The magnitude of the most-positive rate of terminus advance is well correlated with ice speed of the floating ice tongue, as one might expect if a glacier is not calving (Fig. DR2).

ANALYSIS

No strong spatial or temporal pattern of changing glacier length is apparent (Fig. 2). A weak pattern, however, may exist in the northern glaciers, northward of the Borchgrevink Glacier (ID 9). These were observed to advance between ca. 1999 until 2014.

To further evaluate the variations in terminus positions, we investigated three possible controls: ice speed of the ice tongue, calving rate, and the buttressing effect of sea ice. Neither the ice speed of the floating ice tongues nor changes in regional ice-sheet motion showed any spatial or temporal pattern (Table DR4; Figs. DR3–DR5). However, the regional ice speed generally increased everywhere during the 2008–2014 period, by $15–65 \pm 14$ m yr^{-1}. The calving rate, calculated from the difference between rate of change in terminus position and speed of the ice tongue, also showed no spatial or temporal patterns (Fig. DR6). Rates ranged to as much as −1990 m yr^{-1} with an average of −97 m yr^{-1} (Table DR5). The presence of sea ice can buttress the ice front, mechanically stabilizing the glacier face against calving (Reeh et al., 2001; Amundson et al., 2010), and its presence indicates cooler near-surface ocean temperatures which also may reduce melting of the ice tongue. To examine whether the presence of sea ice was important, calving rates were plotted as a function of yearly maximum sea ice area (Markus and Cavalieri, 2000) averaged over the time intervals equal to those of terminus change and calving. No obvious correlations were apparent (Fig. DR7); however, smaller ice tongues within embayments and surrounded by landfast ice may be affected (Miles et al., 2017).

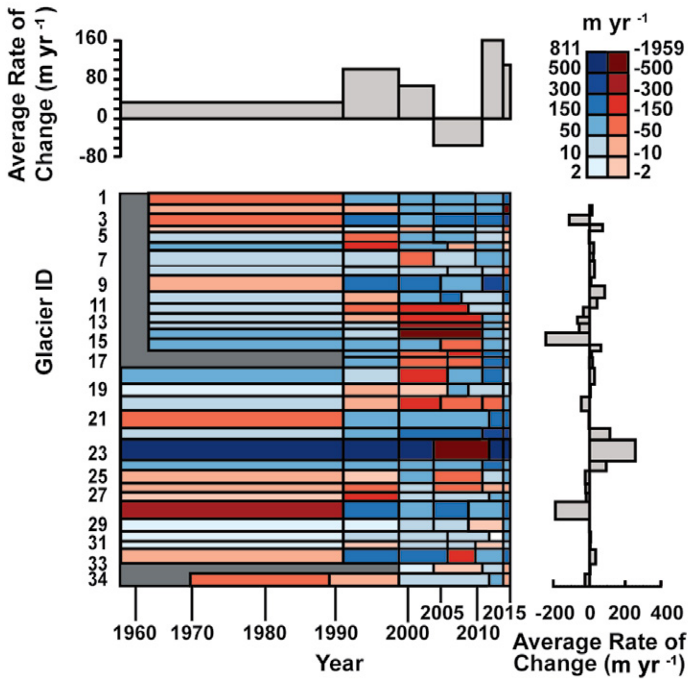

Figure 2. Advance (blue) or retreat (red) rate (m yr^{-1}) of glaciers along west coast of Ross Sea (Antarctica). Time intervals are defined by available maps and imagery. Top graph is average change over time for all glaciers in the study, weighted by glacier width at grounding line. Because time intervals for individual glaciers are not identical, we use periods with the most glacier measurements. Colored graph shows change in glaciers from north (glacier ID 1) to south (ID 34) with time. Row height along *y*-axis is glacier area normalized by area of David Glacier–Drygalski Ice Tongue (ID 23) using logarithmic scale. Gray indicates no data. Graph at lower right is average advance or retreat rate over entire period of measurement for each glacier.

DISCUSSION AND CONCLUSIONS

No strong spatial or temporal patterns of terminus changes, calving rates, or flow speed have emerged from this analysis, supporting the results of previous studies in this region (Frezzotti, 1993, 1997; Miles et al., 2013). The similarity to the results of Frezzotti (1993, 1997) is not exact because he tracked area change whereas we measured length changes. In some cases, Frezzotti (1993) showed area shrinkage during a period of glacier advance due to ice loss along the side of the ice tongue. Since 1959, the net change for all the glaciers, weighted by glacier width at the grounding line, has been advance (Fig. 2). However, this result is dominated by the David Glacier–Drygalski Ice Tongue (Fig. DR3; Table DR1). A weak pattern of glacier advance since the late 1990s appears in the northernmost glaciers and is consistent with an increase in snow accumulation for this region that does not extend further south (Bromwich et al., 2011; Shepherd et al., 2012; Zwally and Giovinetto, 2011). This increase coincides with the start of the cooling trend on the Antarctic Peninsula, but may be a coincidence because the atmospheric forcing in this region is weak compared to that over the Antarctic Peninsula (Turner et al., 2016).

Floating ice tongues that extend beyond the coastline into the open ocean seem to be a feature of Antarctica in the present climate (the ice tongues of Greenland's Petermann and Zachariæ Isstrøm glaciers are in protected fjords or bays). The tongues thin with distance from the grounding zone (the region where the grounded glacier transitions to floatation) due to internal deformation from the force of buoyancy (Cuffey and Paterson, 2010), much like ice shelves. The ice tongues also thin by ablation on both upper and lower surfaces from atmospheric-driven sublimation and melting and by ocean-driven melting, respectively.

The length of a floating ice tongue is likely determined by the flux of ice to the ocean, the internal ice temperature, the overall rate of thinning, and degree of flexure caused by currents, tides, and storms. Initially, the formation of an ice tongue must survive the considerable hinging forces present at the grounding zone (e.g., Holdsworth, 1969). Basal melting at the grounding zone may also be a significant factor (Holdsworth, 1982; Legrésy et al., 2004) in warmer glacier ice and/or warmer deep ocean waters. However, for the cold-cored Drygalski Ice Tongue, Frezzotti (1993) pointed to bottom freezing as a mechanism strengthening the ice and perhaps contributing to its further extension. Calving of the ice results from ice flexure and crack propagation, arising from exposure to the tractive forces in the open ocean. Clearly, thinner ice is more easily fractured, and so a length limit may be set by the steady reduction of ice thickness and the presence of torque forces. The presence of more extended ice tongues in the Ross Sea and their absence in the Weddell Sea (Holdsworth, 1969) may well be a result of cold internal ice from East Antarctica, relatively low torque from winds, and sea ice, in contrast to the Weddell coast of the Antarctic Peninsula with far warmer ice and strong year-round northerly sea ice drift.

Overall, the role of sea ice is unclear. On the one hand, sea ice can mechanically buttress the ice tongue (Holdsworth, 1982) and dampen swell that flexes the ice, but on the other, wind shear across sea ice abutting the glacier tongue can impart a tractive force. The absence of exposed ice tongues in Greenland and the relatively small number elsewhere around Antarctica, compared to the nearly ubiquitous presence of ice shelves along the coastal western Ross Sea, suggests that the set of conditions required to form ice tongues must be uncommon.

No strong latitudinal or temporal pattern in glacial extent is present. At best, the northernmost glaciers are advancing slightly. The stability of these glaciers over the past half century is a stark contrast to the rapidly shrinking glaciers of the Antarctic Peninsula and indicates that the glacial effects of climate change have yet to reach this region of Antarctica.

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REFERENCES CITED

Amundson, J.M., Fahnestock, M., Truffer, M., Brown, J., Luthi, M.P., and Motyka, R.J., 2010, Ice mélange dynamics and implications for terminus stability, Jakobshavn Isbræ, Greenland: Journal of Geophysical Research, v. 115, F01005, doi:10.1029/2009JF001405.

Bindschadler, R., Choi, H., and ASAID Collaborators, 2011, High-resolution image-derived grounding and hydrostatic lines for the Antarctic Ice Sheet (version 1): Boulder, Colorado, National Snow and Ice Data Center Data Set NSIDC-0489, doi:10.7265/N56T0JK2 (accessed April 2015).

Bromwich, D.H., Nicolas, J.P., and Monaghan, A.J., 2011, An assessment of precipitation changes over Antarctica and the Southern Ocean since 1989 in contemporary global reanalysis: Journal of Climate, v. 24, p. 4189–4209, doi:10.1175/2011JCLI4074.1.

Cook, A.J., Fox, A.J., Vaughan, D.G., and Ferrigno, J.G., 2005, Retreating glacier fronts on the Antarctic Peninsula over the past half-century: Science, v. 308, p. 541–544, doi:10.1126/science.1104235.

Cuffey, K.M., and Paterson, W.S.B., 2010, The Physics of Glaciers: Burlington, Massachusetts, Butterworth-Heinemann Elsevier, 704 p.

Fahnestock, M., Scambos, T., Moon, T., Gardner, A., Haran, T., and Klinger, M., 2016, Rapid large-area mapping of ice flow using Landsat 8: Remote Sensing of Environment, v. 185, p. 84–94, doi:10.1016/j.rse.2015.11.023.

Ferrigno, J.G., Foley, K.M., Swithinbank, C., and Williams, R.S., Jr., 2010, Coastal-change and glaciological map of the Ross Island area, Antarctica: 1962–2005: U.S. Geological Survey Geologic Investigations Series Map I-2600-I, 1 map sheet, scale 1:1,000,000, 23 p. text.

Fountain, A.G., Basagic, H.J., and Niebuhr, S., 2016, Glaciers in equilibrium, McMurdo Dry Valleys, Antarctica: Journal of Glaciology, v. 62, p. 976–989, doi:10.1017/jog.2016.86.

Frezzotti, M., 1993, Glaciological study in Terra Nova Bay, Antarctica, inferred from remote sensing analysis: Annals of Glaciology, v. 17, p. 63–71, doi:10.1017/S0260305500012623.

Frezzotti, M., 1997, Ice front fluctuation, iceberg calving flux and mass balance of Victoria Land glaciers: Antarctic Science, v. 9, p. 61–73, doi:10.1017/S0954102097000096.

Frezzotti, M., and Polizzi, M., 2002, 50 years of ice-front changes between the Adélie and Banzare Coasts, East Antarctica: Annals of Glaciology, v. 34, p. 235–240, doi:10.3189/172756402781817897.

Holdsworth, G., 1969, Flexure of a floating ice tongue: Journal of Glaciology, v. 8, p. 385–397, doi:10.1017/S0022143000026976.

Holdsworth, G., 1982, Dynamics of Erebus Glacier Tongue: Annals of Glaciology, v. 3, p. 131–137, doi:10.1017/S0260305500002652.

Kunz, M., King, M.A., Mills, J.P., Miller, P.E., Fox, A.J., Vaughan, D.G., and Marsh, S.H., 2012, Multi-decadal glacier surface lowering in the Antarctic Peninsula: Geophysical Research Letters, v. 39, L19502, doi:10.1029/2012GL052823.

Legrésy, B., Wendt, A., Tabaco, I., Rémy, F., and Dietrich, R., 2004, Influence of tides and tidal current on the Mertz Glacier, Antarctica: Journal of Glaciology, v. 50, p. 427–435, doi:10.3189/172756504781829828.

Liu, H., Jezek, K., Li, B., and Zhao, Z., 2001, Radarsat Antarctic Mapping Project digital elevation model (version 2): Boulder, Colorado, NASA National Snow and Ice Data Center Distributed Active Archive Center, https://nsidc.org/data/docs/daac/nsidc0082_ramp_dem.gd.html (accessed September 2014).

Markus, T., and Cavalieri, D.J., 2000, An enhancement of the NASA Team sea ice algorithm: IEEE Transactions on Geoscience and Remote Sensing, v. 38, p. 1387–1398, doi:10.1109/36.843033.

Miles, B.W.J., Stokes, C.R., Vieli, A., and Cox, N.J., 2013, Rapid climate-driven changes in outlet glaciers on the Pacific coast of East Antarctica: Nature, v. 500, p. 563–566, doi:10.1038/nature12382.

Miles, B.W., Stokes, C.R., and Jamieson, S.S., 2017, Simultaneous disintegration of outlet glaciers in Porpoise Bay (Wilkes Land), East Antarctica, driven by sea ice break-up: The Cryosphere, v. 11, p. 427–442, doi:10.5194/tc-11-427-2017.

Mouginot, J., Scheuchl, B., and Rignot, E., 2017, MEaSUREs Antarctic boundaries for IPY 2007–2009 from satellite radar (version 2): Coastline Antarctica: Boulder, Colorado, NASA National Snow and Ice Data Center Distributed Active Archive Center, doi:10.5067/AXE4121732AD.

Oliva, M., Navarro, F., Hrbáček, F., Hernández, A., Nývlt, D., Pereira, P., Ruiz-Fernández, J., and Trigo, R., 2017, Recent regional climate cooling on the Antarctic Peninsula and associated impacts on the cryosphere: The Science of the Total Environment, v. 580, p. 210–223, doi:10.1016/j.scitotenv.2016.12.030.

Parkinson, C.L., and Cavalieri, D.J., 2012, Antarctic sea ice variability and trends, 1979–2010: The Cryosphere, v. 6, p. 871–880, doi:10.5194/tc-6-871-2012.

Polar Geospatial Center, 2014, Antarctica mosaic 50 cm panchromatic: St. Paul, Minnesota, Polar Geospatial Center, http://viewers.apps.pgc.umn.edu/antarctica/ (accessed July 2014).

Reeh, N., Thomsen, H.H., Higgins, A.K., and Weidick, A., 2001, Sea ice and the stability of north and northeast Greenland floating glaciers: Annals of Glaciology, v. 33, p. 474–480, doi:10.3189/172756401781818554.

Rignot, E., Mouginot, J., and Scheuchl, B., 2011, MEaSUREs InSAR-based Antarctica ice velocity map (version 1): 450 m: Boulder, Colorado, NASA National Snow and Ice Data Center Distributed Active Archive Center, doi:10.5067/D7GK8F5J8M8R (accessed March 2015).

Scambos, T.A., Hulbe, C., Fahnestock, M., and Bohlander, J., 2000, The link between climate warming and the break-up of ice shelves in the Antarctic Peninsula: Journal of Glaciology, v. 46, p. 516–530, doi:10.3189/172756500781833043.

Scambos, T.A., Berthier, E., Haran, T., Shuman, C.A., Cook, A.J., Ligtenberg, S.R.M., and Bohlander, J., 2014, Detailed ice loss pattern in the northern Antarctic Peninsula: Widespread decline driven by ice front retreats: The Cryosphere, v. 8, p. 2135–2145, doi:10.5194/tc-8-2135-2014.

Screen, J.A., and Simmonds, I., 2012, Half-century air temperature change above Antarctica: Observed trends and spatial reconstructions: Journal of Geophysical Research, v. 117, D16108, doi:10.1029/2012JD017885.

Shepherd, A., et al., 2012, A reconciled estimate of ice-sheet mass balance: Science, v. 338, p. 1183–1189, doi:10.1126/science.1228102.

Simmonds, I., 2015, Comparing and contrasting the behaviour of Arctic and Antarctic sea ice over the 35 year period 1979–2013: Annals of Glaciology, v. 56, p. 18–28, doi:10.3189/2015AoG69A909.

Steig, E.J., Schneider, D.P., Rutherford, S.D., Mann, M.E., Comiso, J.C., and Shindell, D.T., 2009, Warming of the Antarctic ice-sheet surface since the 1957 International Geophysical Year: Nature, v. 457, p. 459–462, doi:10.1038/nature07669.

Turner, J., Colwell, S.R., Marshall, G.J., Lachlan-Cope, T.A., Carleton, A.M., Jones, P.D., Lagun, V., Reid, P.A., and Iagovkina, S., 2005, Antarctic climate change during the last 50 years: International Journal of Climatology, v. 25, p. 279–294, doi:10.1002/joc.1130.

Turner, J., Lu, H., White, I., King, J.C., Phillips, T., Hosking, J.S., Bracegirdle, T.J., Marshall, G.J., Mulvaney, R., and Deb, P., 2016, Absence of 21st century warming on Antarctic Peninsula consistent with natural variability: Nature, v. 535, p. 411–415, doi:10.1038/nature18645.

U.S. Geological Survey, 1972, Ross Island, Antarctica [map]: Washington, D.C.: United States Department of the Interior, U.S. Geological Survey Topographic Reconnaissance Series, scale 1:250,000.

U.S. Geological Survey, 1970, Cape Adare, Antarctica [map] (compiled 1968): Washington, D.C.: United States Department of the Interior, U.S. Geological Survey Topographic Reconnaissance Series, scale 1:250,000.

U.S. Geological Survey, 1970, Cape Hallett, Antarctica [map] (compiled 1968): Washington, D.C.: United States Department of the Interior, U.S. Geological Survey Topographic Reconnaissance Series, scale 1:250,000.

U.S. Geological Survey, 1969, Coulman Island, Antarctica [map] (compiled 1967): Washington, D.C.: United States Department of the Interior, U.S. Geological Survey Topographic Reconnaissance Series, scale 1:250,000.

U.S. Geological Survey. Mount Melbourne, Antarctica [map]. Compiled 1967. 1:250,000. Topographic Reconnaissance Series. Washington, D.C.: United States Department of the Interior, USGS, 1968.

U.S. Geological Survey, 1968, Relief Inlet, Antarctica [map] (compiled 1966): Washington, D.C.: United States Department of the Interior, U.S. Geological Survey Topographic Reconnaissance Series, scale 1:250,000.

U.S. Geological Survey, 1965, Franklin Island, Antarctica [map] (compiled 1962): Washington, D.C.: United States Department of the Interior, U.S. Geological Survey Topographic Reconnaissance Series, scale 1:250,000.

Zwally, H.J., and Giovinetto, M.B., 2011, Overview and assessment of Antarctic ice-sheet mass balance estimates: 1992–2009: Surveys in Geophysics, v. 32, p. 351–376, doi:10.1007/s10712-011-9123-5.

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