Late Holocene Relative Sea-Level Reconstruction near Palmer Station, northern Antarctic Peninsula strongly controlled by Late Holocene ice mass changes

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Abstract

Many studies of Holocene relative sea-level (RSL) changes across Antarctica assume that their reconstructions record uplift from glacial isostatic adjustment caused by the demise of the Last Glacial Maximum (LGM) ice sheets. However, recent analysis of GPS observations suggests that mantle viscosity beneath the Antarctic Peninsula is weaker than previously thought, which would imply that solid Earth motion is not controlled by post-LGM ice-sheet retreat but instead by late Holocene ice-mass changes. If this hypothesis is correct, one might expect to find Holocene RSL records that do not reflect a monotonic decrease in the rate of RSL fall but show variations in the rate of RSL change through the Holocene. We present a new record of late Holocene RSL change from Torgersen Island near Palmer

Station in the western Antarctic Peninsula that shows an increase in the rate of relative sea-level fall from 3.0±1.2 mm/yr to 5.1±1.8 mm/yr during the late Holocene. Independent studies of the glacial history of the region provide evidence of ice-sheet changes over similar time scales that may be driving this change. When our RSL records are corrected for sea-surface height changes associated with glacial isostatic adjustment (GIA), the rate of post-0.79 ka land uplift at Torgersen Island, 5.3±1.8 mm/yr, is much higher than the rate of uplift recorded at a nearby GPS site at Palmer Station prior to the Larsen B breakup in 2002 AD (1998-2002 AD; <0.1 mm/yr), but similar to the rates observed after 2002 AD (2002-2013 AD; 6-9 mm/yr). This substantial variation in uplift rates further supports the hypothesis that Holocene RSL rates of change are recording responses to late Holocene and recent changes in local ice loading rather than a post-LGM signal across portions of the Antarctic Peninsula. Thus Middle-to-Late Holocene RSL data may not be an effective tool for constraining the size of the LGM ice sheet across portions of the Antarctic Peninsula underlain by weaker mantle. Current global-scale GIA models are unable to predict our observed changes in Late Holocene RSL. Complexities in Earth structure and neoglacial history need to be taken into consideration in GIA models used for correcting modern satellite-based observations of ice-mass loss.

1. Introduction

As early as Nichols (1960), the nature of relative sea-level (RSL) change across Antarctica was interpreted to reflect glacial isostatic adjustment (GIA) in response to the decay of the LGM ice sheets. Initial attempts at fitting observations of RSL change to model predictions of GIA in Antarctica used a relatively strong Earth rheology (Pallas et al., 1997; Zwartz et al., 1998; Bassett et al., 2007). In addition, many of these initial GIA studies assumed that the ice sheet experienced a continuous demise with few if any readvances (and subsequent retreats) through the Holocene (e.g. ANT3, Nakada and Lambeck, 1988; ICE-5G, Peltier, 2004; W12, Whitehouse et al., 2012b). These initial GIA model predictions (Pallas et al., 1997; Zwartz et al., 1998; Bassett et al., 2007; Whitehouse et al., 2012b) appear to fit the sparse RSL observations available from the continent and its neighboring islands (Nichols, 1960; Pallas et al., 1997; Hall and Denton, 1999; Baroni and Hall, 2004; Roberts et al., 2011; Simkins et al., 2013b). However, both the existence of a strong Earth rheology and the assumption of no Holocene ice readvances have been challenged by recent studies showing local ice advances in parts of the northern Antarctic Peninsula (Smith, 1982; Hansom and Flint, 1989; Hjort et al., 1997; Hall, 2007; Hall et al., 2010).

Recent warming across the Antarctica Peninsula and Southern Ocean has brought about a natural experiment in which to test the existence of a relatively stiff and strong rheology beneath the Antarctic Peninsula. Remote sensing along with historical records have been used to reconstruct the amount of ice-mass loss over historical time periods (Nield et al., 2012; 2014; Zhao et al., 2017). These reconstructions coupled with GPS observations of uplift effectively constrain the Earth structure beneath the Antarctic Peninsula (Nield et al., 2014; Zhao et al., 2017). In particular, Nield et al. (2014) show that the large increase in uplift rates recorded in the GPS data since ice-shelf collapse in 2002 cannot be explained by elastic deformation alone but reflect a visco-elastic Earth response. Furthermore, the amount of visco-elastic response suggests that some regions of the Antarctic Peninsula are underlain by a much weaker rheology than early and global GIA models have suggested (Ivins et al., 2011; Nield et al., 2014).

Additional work focused on reconstructing the history of the Antarctic Peninsula Ice Sheet has also provided evidence for late Holocene ice-sheet oscillations (Hjort et al., 1997; Hall et al., 2010). In addition, GPS observations in areas far removed from ice shelves undergoing recent collapse have found that post-LGM ice loss alone cannot explain current rates of rebound but require late Holocene ice-mass changes (Bradley et al., 2015; Wolstencroft et al., 2015). Only recently have GIA-models considered the possibility of Holocene ice-sheet oscillations (Ivins et al., 2011). Such oscillations could result in observable changes in the rate of RSL change through the Holocene if portions of the Antarctic Peninsula are underlain by a relatively weak Earth structure. Such changes in the rate of RSL fall have been documented in the South Shetland Islands (Hall, 2010; Watcham et al., 2011; Simms et al., 2012), which might be expected to have a weaker rheology given their location overlying an active subduction zone, but have yet to be documented for the Antarctic Peninsula proper.

The purpose of this study is to use optically-stimulated luminescence (OSL) ages of raised beaches to reconstruct a new RSL record from the northern Antarctic Peninsula continental margin in an area where prior GPS work has documented rapid uplift suggestive of a relatively weak rheology. If the Antarctic Peninsula is underlain by a weaker-than-average Earth structure and its ice sheet has been subject to Holocene oscillations, one would expect to find evidence for variable rates of RSL change during the late Holocene rather than the monotonic decrease predicted by simple post-LGM ice retreat. We test this hypothesis. Furthermore, after correcting our new RSL record for GIA-induced changes in the sea-surface height and assuming minimal influences from steric impacts on the Holocene record of sea-level change, we compare our new rate of Holocene uplift to recent GPS (1998-present) measurements of uplift at Palmer Station along the northern Antarctic Peninsula. This comparison is used to determine if recent rates of RSL fall are unprecedented in the late Holocene.

2. Background

2.1 Late Holocene Sea Levels and Rates of RSL Change

One of the best millennial-scale archives of glacial isostatic adjustment is Holocene RSL records. Relative sea-level reconstructions for the western Antarctic Peninsula are available for the South Shetland Islands at the northern tip of the Antarctic Peninsula (John and Sugden, 1971; Bentley et al., 2005; Hall, 2010; Watcham et al., 2011; Simms et al., 2012) and in Marguerite Bay to the south (Nichols, 1960; Bentley et al., 2005; Hodgson et al., 2013; Simkins et al., 2013b), leaving a 600+ km stretch of coast with little to no constraints on Holocene sea-level changes (Fig. 1). Along the eastern Antarctic Peninsula the only sea-level indicators are found at Beak Island (Roberts et al., 2011) and James Ross Island (JRI) (Ingolfsson et al., 1992; Hjort et al., 1997) (Fig. 1).

Relative sea-level reconstructions from the South Shetland Islands record a recent sea-level fall on the order of 6 m within the last 300-500 years at a rate possibly as high as 12.5 mm/yr preceded by an overall 10-12 m fall in sea level since 6 ka (Bentley et al., 2005; Hall, 2010; Watcham et al., 2011; Simms et al., 2012). The rapid RSL changes within the South Shetland Islands (Hall, 2010; Watcham et al., 2011) have been attributed to the relatively weak rheology beneath the active arc (Simms et al., 2012). Sea-level records from Marguerite Bay show a fall in sea level from an approximate highstand of 20-22 m at 7-7.5 ka to <5 m around 2-2.5 ka (Bentley et al., 2005; Hodgson et al., 2013; Simkins et al., 2013b). This fall may have been preceded by a rise in RSL contributing to the formation of a prominent scarp within northern Marguerite Bay (Simkins et al., 2013b). Since 2-2.5 ka, sea levels within Marguerite Bay have fallen at a rate of less than 1.4 mm/yr (Simkins et al., 2013b).

The records from the eastern Antarctic Peninsula are sparser, with three sea-level index points from Beak Island (Roberts et al., 2011) and two sets of marine/beach deposits from northern JRI (Ingolfsson et al., 1992; Hjort et al., 1997) (Fig. 1). The age-elevation relationships of the three indices from Beak Island are precisely established from isolation basins (Roberts et al., 2012). This work suggests rates of RSL fall decreased from 3.9 mm/yr between 8 ka and 6.9 ka to 2.1 mm/yr between 6.9 ka and 2.9 ka, leveled off to 1.6 mm/yr between 2.9 ka and 1.8 ka, and further decreased to 0.29 mm/yr over the last 1.8 ka. Constraints on rates of RSL changes from the two sets of raised marine features found on JRI (Hjort et al., 1992) are not as clear. At the first site, Brandy Bay, two sets of raised beaches are found, with a higher, more weathered set of beaches at elevations from 30-18 m and a lower, better preserved, set of beaches at less than 16 m (Ingolfsson et al., 1992; Hjort et al., 1997). At a second site on JRI, The Naze, the lower set of beaches is found at elevations less than 18 m (Hjort et al., 1997). In addition to the beaches at The Naze, cliff exposures reveal an 18 m thick succession of late Holocene glacial-marine and shoreface deposits with *in situ* mollusks. The succession places sea levels higher than 18 m around 7.5 ka (Hjort et al., 1997). Hjort et al. (1997) therefore assign the older beaches with an upper limit of 30 m to an age of 7.5 ka and the lower beaches with an upper limit of 16-18 m to an age of 4.2-4.5 ka based on ages from *in situ* mollusks from another lower section of glacial marine deposits close to the lower raised beaches within Brandy Bay. The stratigraphy and correlations provide a well-documented framework for Holocene sea-level changes but the elevation-age relationship cannot be directly confirmed.

2.2 Recent Changes across the Antarctic Peninsula

The Antarctic Peninsula is one of the most rapidly warming places on the planet (Hansen et al., 1999; Vaughan et al., 2003). Recent warming has brought dramatic changes to the region including an increase in ice melting recorded in ice core data (Abram et al., 2013), the disappearance of ice shelves (Vaughan and Doake, 1996; Rott et al., 1996; Hodgson et. al., 2011), changes in ocean circulation (Jourdain et al., 2017), the retreat of tidewater glaciers (Cook et al., 2005) and sea ice (Cavalieri and Parkinson, 2008; Jourdain et al., 2017), and changes in regional ecology and ecosystems (Moline et al., 2008; Montes-Hugo et al., 2009; Mintenbeck and Torres, 2017; Amesbury et al., 2017). The retreat of glaciers due to both warming and the breakup of ice shelves reflects an increase in the regional rate of mass loss (Rignot et al., 2008). This decrease in mass was accompanied by an increase in uplift rates recorded in GPS observations across the Antarctic Peninsula (Thomas et al., 2011; Nield et al., 2014). This acceleration in uplift is more pronounced in the northern Antarctic Peninsula than the central Antarctica Peninsula (Thomas et al., 2011; Nield et al., 2014). For example, GPS observations from Palmer Station record a 110-fold increase in uplift rates following the demise of the Larsen B Ice Shelf in 2002 (Thomas et al., 2011; Nield et al., 2014) while farther south along the Antarctic Peninsula at Rothera Station, uplift rates increased by a factor of less than 5 (Thomas et al., 2011). Part of these differences can be attributed to differences in the local magnitude of grounded ice loss but part has been attributed to differences in the rheology of the underlying earth (Zhao et al., 2017).

2.3 Role of other drivers of sea-level change

In addition to GIA, rock uplift, and ocean volume changes, steric changes such as changing ocean temperatures and changes in wind patterns can also drive relative sea-level changes. Over the last 50 years, ocean temperatures have warmed nearly 1°C within the Bellingshausen and Amundsen Seas (Meredith and King, 2005). This warming along the Antarctic Peninsula is seen to be a response to global climate variability including changes in the Atlantic Multidecadal Oscillation (Li et al., 2014). Sea-surface temperatures across the Antarctic Peninsula have also changed throughout the late Holocene with the amplitude of oscillations reaching as much as 1.8°C over the last 2 ka (Shevenell et al., 2011). However, such temperature changes are unlikely to have caused sea-level variations of more than a few tens of cm (Landerer et al., 2007). For example, along the Pacific coasts of North and South America, local ocean temperature increases of up to 2-5°C during El Nino years brought about only 20 to 30 cm of sea-level change (Hamlington et al., 2015). Some of these same studies, although not specifically focused on the western Antarctic Peninsula, only show a few cm's of sea-level variability across the Southern Ocean near the western Antarctica Peninsula (Hamlington et al., 2015). As for wind changes, based on the analysis of tide-gauge data at Vernadsky Station approximately 50 km south of Palmer Station, Aoki (2002) found less than 10 cm of sea-level variability due to changes in winds associated with the Southern Hemisphere Annular Mode, with a mean increase in wind speeds of 1 m/s increasing sea levels by only 1 cm.

*2.4 Evidence for late Holocene ice-changes across the western Antarctic Peninsu*la

Evidence of late-Holocene advances and retreats within the western Antarctic Peninsula are found both in the terrestrial and marine record. Terrestrial records of glacial/ice sheet changes near our field site on Torgerson Island include observations of floral development on the rock-exposed headlands near Palmer Station (Smith, 1982) as well as reworked material within tills (Hall et al., 2010). Although no terminal moraines have been identified, a distinct decrease in the development and changes in the types of mosses and lichens suggest the ice extended approximately 150 m farther seaward of its 1977 AD limits (the ice has continued to retreat since 1977) within the last millennium (Smith, 1982). The remnants of what appeared to be moss overridden by the ice dated to 0.54 ka (0.67 ka to 0.32 ka) suggesting the ice advanced sometime after this date (average of 3 ages recalibrated from Smith, 1982). In addition to this advance, Hall et al. (2010) found evidence of a more landward position (less ice) of the glacier near Palmer Station within the late Holocene. Mosses incorporated into the modern moraines near Palmer Station suggest ice-free areas existed beneath the present glacier around 0.7-0.97 ka indicating a retreat of the glacier prior to its advance suggested by Smith (1982). Furthermore, other mosses within the banks point to other potential periods of less extensive ice around 3.7 and 5.6 ka (Hall et al., 2010).

Outside of the region around our study area, terrestrial evidence for late Holocene ice-sheet oscillations have been found on the South Shetland Islands (Curl, 1980; Clapperton and Sugden, 1988; Birkenmajer, 1998; Hall, 2007; Simms et al., 2012), Brabant Island (Hansom and Flint, 1989), and James Ross Island (Hjort et al., 1997). Of the records on the South Shetland Islands, the best dated is probably that of Hall (2007) who obtained 28 radiocarbon ages within a well-developed moraine on King George Island. The ages suggest the moraine formed from an advance less than 650 years ago that was preceded by less extensive ice present sometime after ~3.5 ka (Hall, 2007). Hansom and Flint (1989) and Hjort et al. (1997) both dated reworked mollusk shells within moraines suggesting an advance sometime around 5 ka.

Marine records from the fjords across the western Antarctic Peninsula also point to glacial and climatic fluctuations throughout the late Holocene. Direct records of an ice advance leading up to the Little Ice Age at approximately 0.2 ka are recorded in cores from Barilari Bay approximately 130 km south of Palmer Station (Christ et al., 2015). Proxy records from the Palmer Deep 25 km south of Palmer Station point to almost 1.3°C warming of sea-surface temperatures between 1.8 ka and 1.6 ka with a similar magnitude cooling between 0.5 ka and 0.1 ka with no fewer than 8 additional but smaller amplitude sea-surface temperature changes since 1.8 ka (Shevenell et al., 2011). Several other proxy records of climatic changes have been retrieved from marine sediments within the Palmer Deep (Leventer et al., 1996; Shevenell and Kennett, 2002) and elsewhere across the western Antarctic Peninsula (Heroy et al., 2008; Allen et al., 2010; Yoon et al., 2010; Reilly et al., 2016; Kim et al., 2018) all pointing to climatic and potentially glacial fluctuations through the late Holocene.

2.5 *Field Site*

Our new RSL site at Torgersen Island is located along the western Antarctic Peninsula within the Palmer Archipelago on the southern side of Anvers Island (Figs. 1 and 2). It is 285 km south and 375 km north of the closest RSL records for the western Antarctica Peninsula in the South Shetland Islands and Marguerite Bay, respectively. It is located less than 1 km to the west of the Palmer Station GPS site, which itself lies on a small peninsula extending from Anvers Island (Fig. 2). The region consists of isolated rock drumlins and other glacially-carved landforms covered in a veneer of late Holocene till, outwash, and raised marine features (Hall et al., 2010). Anvers Island partly shields the Torgersen Island region from the dominant swell originating in the southern Pacific Ocean and traveling through the Bellingshausen Sea (Fig. 1).

Although Torgersen Island is located on the Bellingshausen Sea, the region is only separated from the Weddell Sea and the former location of the Larsen B Ice Shelf by a narrow spine of mountains comprising the northern Antarctic Peninsula (Fig. 1). As such it is located less than 100 km from the former Larsen B Ice Shelf and within the influence of increased uplift rate due to ice flow acceleration following the recent demise of the ice shelf (Thomas et al., 2011; Nield et al., 2014).

3. Methods

3.1 Basics of OSL dating

Optically stimulated luminescence (OSL) dating determines the last exposure of mineral grains to sunlight. The age range is from recent decades to 200,000 years ago with an age-uncertainty of 5-10%. New methodologies promise an extension of the upper age limit to 1 million years (e.g. Porat et al., 2009). Bøtter-Jensen et al. (2003), a review by Rhodes (2011), and a series of dedicated contributions in the recently published Encyclopedia of Scientific Dating Methods (Rink and Thompson, 2013) give a detailed description of the OSL method. OSL dating works on the principle that radiation – from U, Th, K, and from cosmic rays – ionizes atoms within silicate minerals like quartz and feldspar. The freed electrons become trapped at light sensitive crystal defects. When exposed to sunlight, the electrons are released from the traps. In returning to their original states they emit luminescence and the mineral is reset. Upon burial trapped electrons accumulate again, and their number is proportional to the burial time and the radiation exposure, often termed the “dose”. The rate of irradiation, the “dose rate,” can be calculated from the cosmic flux as well as the U, Th, and ^{40}K concentrations of the surrounding materials. The OSL signal is proportional to the dose and can be measured by exposing the mineral to light in a controlled setting. An age since burial can be determined by dividing the dose by the dose rate. The lower age limit of the methodology is determined by the smallest measurable signal. The upper age limit depends on the saturation range of the sample, where the signal shows only little increase with dose. Typically this procedure is conducted on sand-sized quartz grains (Murray and Wintle, 2003), but recent advances have allowed for the use of OSL on cobble surfaces (Simms et al., 2011; Simkins et al., 2016).

3.2 OSL Sampling

Torgersen Island as well as nearby Litchfield Island and Biscoe Point were visited in the austral fall of 2010 (Figs. 1 and 2). The deposits from Torgersen and Litchfield Islands were described and surveyed using a survey-grade Trimble GPS. However, the samples from Biscoe Point were surveyed using an altimeter due to difficulties associated with the GPS equipment in the field. Elevations were initially calculated relative to the water level elevation at the time of the survey. The offset between the water level elevation at the time of the survey and mean sea level was corrected for using the tide gauge located at Palmer Station. Cobbles for optically stimulated luminescence (OSL) dating were collected under a light-proof tarpaulin to avoid sunlight exposure (Simms et al., 2011). Samples were taken from raised beaches at 5.0 and 7.2 m above modern sea levels (Simkins et al., 2015) as well as the modern beaches on nearby Litchfield Island (0.5 m above sea level) and Biscoe Point (~2.0 m above sea level). The latter two sites were sampled in order to determine if modern beaches gave an OSL age of zero.

3.3 OSL Ages

In the laboratory the outside 1 mm of the undersides of the cobbles was isolated using a Buehler Isomet 1000 precision saw. These thin rock slices were crushed lightly using a ceramic mortar and pestle to disaggregate the constituent mineral crystals (Simkins et al., 2016). The crystal segregates were sieved to isolate grains between 63 µm and 250 µm. Following treatment with 3.75% HCl and 27% H_2O_2, the isolated grains underwent density separations of 2.62 and 2.75 g/cm^3 to extract the quartz fraction. The quartz fraction was further etched with 48% HF for 40 minutes. Approximately 100-200 grains (3 mm aliquots) were prepared on sample carriers using a silicone oil spray. Sample carriers were cleaned as described by Simkins et al. (2013a) and selected for negligible intrinsic signals.

The OSL measurements were carried out using a Risø TL/OSL-DA-15 Reader with a built-in 90Sr/90Y beta source. Blue LEDs (470 nm, 31 mW/cm^2) and IR LEDs (875 nm, 110 mW/cm^2) were used for optical stimulation and infrared stimulation, respectively. OSL signals were detected in the UV-window (Hoya U340, 7.5 mm, 340 nm peak) with 1-s counting intervals. Heating rate used was 5°C/s. The single-aliquot regenerative dose (SAR) procedure was used for determining the equivalent doses (Murray and Wintle, 2000; Wintle and Murray, 2006) with high-temperature stimulation (Murray and Wintle, 2003) and a post-IR blue sequence (Wallinga, 2002; Duller, 2003) following the measurement protocol outlined in Simkins et al. (2013a, b). Aliquots for dose calculations were selected according to the reliability test recommended by Wintle and Murray (2006) and described in detail by Simkins et al. (2015). The extracted quartz was dim and aliquots were considered to be reliable if their recuperation was <10%, recycling ratios <20%, and dose deviation <25%. We used the common age model (Galbraith et al., 1999) for final age determinations.

3.4 GIA Modeling

Relative sea-level change is not only a function of vertical land motion (as measured by GPS) but also a function of changes in the height of the sea surface. We used a global GIA model to estimate the change in sea surface height (SSH) attributed to changes in water volume and the gravitational components of GIA in order to isolate what portion of the RSL signal was due to local land motion assumed to be controlled by ice mass changes, without explicitly modelling those ice-mass changes. The sensitivity of the SSH signal to local ice-mass change is discussed below. Sea surface heights can also be influenced by steric (water-temperature) changes and wind-driven stresses. However, as steric and wind-driven changes are only on the order of a few tens of cm’s (Aoki, 2002; Landerer et al., 2007;

McKay et al., 2011; Hamlington et al., 2015), we assumed they were less than a few 10's of cm over the time scales of our RSL reconstructions.

We simulated the GIA process following the methods of Whitehouse et al. (2012b), using the ice model of Whitehouse et al. (2012a) without the Antarctic Peninsula adjustment (Whitehouse et al., 2012b) for Antarctica and ICE-5G for the Northern Hemisphere (Pelter, 2004). In order to determine what impact a local ice-sheet oscillation may have had on the SSH, we also considered three modified ice-sheet models. These modified ice-sheet models were constructed by decreasing the amount of ice within the original Whitehouse et al. (2012a) model to 90% of its original thickness at 1 ka and increasing it to 101%, 105%, and 110% of its original thickness at 0.5 ka. These modifications were applied to the Antarctic Peninsula ice sheet north of 68.5° S latitude. Our GIA model predictions account for the impacts of migrating shorelines, marine-based ice, and rotational feedback (Kendall et al., 2005). We assumed a spherically symmetric, self-gravitating Maxwell viscoelastic Earth structure, but in order to reflect suggestions that different regions of Antarctica may be underlain by different Earth rheology (van der Wal et al., 2015) we considered 24 different Earth models with lithospheric thicknesses of 46, 71, 96, and 120 km, upper mantle viscosities of 0.05, 0.08, 0.1, 0.2, 0.3, 0.5, and 0.8 x 10^{21} Pa s, and lower mantle viscosities of 3, 10, and 20 x 10^{21} Pa s. Our GIA model has an inherent resolution or truncation degree of 256. Lower mantle viscosities of 3 and 20 x 10^{21} Pa s were only used in conjunction with the Earth model that has a lithospheric thickness of 96 km and upper mantle viscosity of 0.3 x 10^{21} Pa s. Although the upper mantle viscosities used are slightly higher (10^{19} versus 10^{18} Pa s) than those suggested by Nield et al. (2014), our selection of upper mantle viscosities spanning 2 orders of magnitude result in less than 0.2 mm/yr of difference to our predictions of sea surface height change over the last 2 ka (see below).

The GIA model allows us to estimate GIA-driven changes to sea surface height, and the deformation of the sea floor at Torgersen Island, during the mid-to-late Holocene. Although the magnitude of solid-Earth deformation is highly sensitive to the Earth model, the sea surface component is relatively insensitive to both the choice of Earth model and the potential local ice sheet oscillations (see results). We therefore use the mean of the 96 model predictions (24 Earth models x 4 ice models) to correct the RSL observations for sea-surface height change, yielding a data-driven estimate for solid-earth deformation during the late Holocene, which may be directly compared with GPS-derived measurements of contemporary deformation. Two standard deviations (rounded up) of the 96 model predictions is used to quantify the error associated with model uncertainties.

4. Results

4.1 Ages

The ages from Torgersen Island presented here were previously reported as part of a compilation of Antarctic raised beach ages (Simkins et al., 2015). The samples from Litchfield Island and Biscoe Point at elevations of 0.5 m and ~2.0 m, respectively, contained no natural dose, suggesting recent exposure and a "zero" age for the modern beach (Table 1). Two ages were obtained from the Torgersen Island beach at 7.2 m (Table 1). These ages are 1480 and 2180 years with 2-sigma errors of 200 and 680 years, respectively. Although they fall within error of one another, the age with the largest error bar, due to high aliquot measurement scatter (i.e. overdispersion), is older by 700 years. We therefore use a weighted mean based on their errors (Taylor, 1997) to determine an age of 1540±190 years for our RSL calculations. The single age from the 5.0 m beach ridge on Torgersen Island was 790 years with a 2-sigma error of 180 years.

4.2 Rates of Holocene Sea-Level Change

The two beach ridges from Torgersen Island have the same geomorphic expression – low amplitude ridge fronting a broad depositional flat bench (Fig. 2). Following Fretwell et al. (2010), we therefore assume that they formed at similar elevations with respect to sea level at the time of their formation. Although no ground-penetrating radar is available from this site, the morphology of a flat bench is suggestive of a depositional plain similar to the strandplain deposits reported from the South Shetland Islands by Lindhorst and Schutter (2014) rather than a storm-built berm (Butler et al., 1999). Thus for calculating rates of RSL change between the formation ages of these two beach ridges, we use the following expression:

$$dRSL/dt = (E_{br1} - E_{br2})/(T_{br1} - T_{br2}) \quad (1)$$

where E_{br1} and E_{br2} are the elevation of the beach ridges in meters and T_{br1} and T_{br2} are their ages in years. The error was determined using the following expression:

$$E = (1/(T_{br1}-T_{br2}))*(\delta E_{br1}\text{^}2+ \delta E_{br1}\text{^}2+(dRSL/dt)\text{^}2 \times (\delta T_{br1}\text{^}2+ \delta T_{br1}\text{^}2))\text{^}0.5 \quad (2)$$

where δE and δT are the errors in the elevations and ages of the beach ridges. Beach-ridge elevation errors were added in quadrature and accounted for errors in the GPS (<0.1 m), 2 standard deviations of the average beach ridge elevations surveyed along strike (<0.3 m), and 2 standard deviations of the average tidal level during the time of the survey (<0.1 m) for a total error of 0.3 m. The rate of RSL fall between the 7.2 and 5.0 m beaches is 3.0±1.2 mm/yr (Table 2).

We were not able to survey the elevation of the modern beach on Torgersen at the time of our OSL collection. However, a recent topographic survey and published map places the modern beach at around 1 m and well below 2 m (Lorenz and Harris, 2014). In addition, we did survey the elevations of modern beaches on nearby Litchfield Island and Biscoe Point, both of which produced OSL ages of 0 or modern (Table 1). The modern beach on Litchfield Island is found at an elevation of 0.5 m (Table 1), while the modern beach surveyed on nearby Biscoe Point is found at an elevation of ~2.0 m (based on altimeter not GPS). These elevations are similar to surveyed modern beaches in other parts of the Antarctic Peninsula, which are found at elevations of less than 2 to 3 meters (Bentley et al., 2005; Hall, 2010; Simms et al., 2011; Fretwell et al., 2011). Their variability likely represents local differences in fetch, grain size, and shallow water bathymetry – all of which impact beach heights. We therefore assign an elevation of the modern-equivalent beach on Torgersen Island to 1.0±1.0 m. Using Eqs. 1 and 2 from above yields a rate of RSL fall of 5.1±1.8 mm/yr since 0.79 ka (Table 2).

4.3 GIA-Induced Sea-Surface Corrections

Our Holocene estimates of rates of relative sea-level change cannot be directly compared with the GPS uplift rates because RSL is a function of more than simply the amount of vertical land motion. However, using a GIA model we estimate the component of RSL change associated with changes in the sea-surface height, and by correcting for this minimize the expected differences between the two types of measurements. Depending on the Earth model, the deglaciation model of Whitehouse et al. (2012a) yields a sea-surface (i.e. non-solid earth) rate component of between 0.38 and 0.21 mm/yr for the time period between 1.54 ka and 0.79 ka and between 0.22 and 0.04 mm/yr for the time period from 0.79 ka to the present, where the positive values indicate gradual SSH rise over these periods (Figs. 3A and 3B). Accounting for possible ice-sheet oscillations over the last 2 ka increases those SSH rates of change by less than 0.02 mm/yr for the time period between 1.54 and 0.79 ka and between 0.01 and -0.03 mm/yr (negative values denote a sea-surface fall) for the time period between 0.79 and the present (Figs. 3C and 3D; the value of -0.03 mm/yr although not shown on Fig. 3D occurs within the predictions using the ice-sheet model with only a 1% increase in ice at 0.5 ka). Combining these potential variations in the SSH results in a total SSH correction of 0.3±0.1 and 0.2±0.1 mm/yr for the time periods of 1.54 to 0.79 ka and 0.79 ka to the present, respectively. Correcting our RSL estimates for GIA-induced SSH changes yields uplift rates of 3.3±1.2 mm/yr between 0.79 ka and 1.54 ka and 5.3±1.8 mm/yr since 0.79 ka and (Table 2; Fig. 4).

Prior to 8 ka, GIA-induced sea-surface height changes are predicted to have kept pace with uplift rates for several millennia (Fig. 5), reflecting the influence of meltwater input from the decaying northern hemisphere ice sheets at this time. However, during the late Holocene, rates of GIA-induced sea-surface height change are predicted to have been minor in comparison to the rates of RSL fall over the same period and were never greater than 0.5 mm/yr over the last 2.5 ka (Table 2, Fig. 5). Predictions for sea-surface height changes vary little across the Antarctica Peninsula (not shown) and thus variations in the rate of RSL fall largely reflect differences in the rate of solid earth deformation.

5.0 Discussion

5.1 Changes in Holocene Rates of RSL Fall

Although the rates of RSL fall derived from the 7.2- to 5.0-m beaches and the 5.0-m to modern beach ridges overlap within error, the central values suggest an increase in the rate of RSL fall by 2.1 ±2.2 mm/yr after 0.79 ka (Table 2). This increase becomes marginally smaller after correcting for sea-surface height changes to obtain uplift rates (2.0±2.2 mm/yr). The sign of this change is opposite to the decrease in rate that would be expected if the late Holocene record of RSL and uplift was dominated by post-LGM isostatic rebound alone. The potential increase may represent the solid-earth response to the late Holocene ice retreat to smaller-than-present margins that is documented for the Marr Ice Piedmont, which lies within 2 km of our beach sites (Hall et al. 2010). Hall et al. (2010) found reworked moss dating to 0.7-0.97 ka along with reworked marine shells within a moraine recently exposed due to recent warming in the Antarctic Peninsula (Hall et al., 2010). This episode of late Holocene ice loss may have been in response to warmer sea-surface temperatures recorded in the region (Shevenell et al., 2011). However, more RSL data points are needed to constrain the exact timing of the RSL rate changes. Such sensitivity of the solid earth to late Holocene glacial retreats would support assertions of a relatively weak underlying upper mantle (Nield et al., 2012; 2014).

A Late Holocene retreat to smaller-than-present margins had to be followed by an ice advance to the pre-1950 glacial margins sometime within the last 800 years. Smith (1982) provides evidence in the form of the density and types of mosses and lichens present near Palmer Station of a glacial advance sometime after 0.54 ka. Such an advance may have dampened any ongoing solid-earth response to the 0.7-0.97 ka retreat documented by Hall et al. (2010); which is weakly shown in our GIA modeling (Fig. 5). Indeed, Nield et al. (2014) suggest that initiation of the viscoelastic response to local ice mass change may be within a few months in this region, leading us to hypothesize that RSL change (largely reflecting changes in uplift rate) during the late Holocene may have been much more variable than is reflected by the time-averaged RSL rates derived from the Holocene beach ridges on Torgersen Island. In particular, rates of RSL fall may have been instantaneously greater than those recorded by the beach ridges, which tend to provide averages for periods during the late Holocene.

5.2 RSL versus GPS Uplift Rates

The late Holocene average uplift rates from 0.79 ka to the present at Torgersen Island, derived by correcting observed RSL rates for the signal due to sea surface height change, are approximately 5 mm/yr greater than the pre-2002 GPS-observed uplift rates measured at Palmer Station (1998-2002; 0.08±1.87 mm/yr; Thomas et al., 2011), and only slightly lower than the rates measured since the breakup of the Larsen B Ice Shelf in 2002 (6.6±2.1 mm/yr based on data from 2002-2013 by Nield et al., 2014; 8.75±0.64 mm/yr based on data from 2002-2010 by Thomas et al., 2011). Nield et al. (2014) demonstrated that the rapid change in GPS-observed uplift rates between pre- and post-2002 could not be explained by an elastic-only response to local ice loss. The authors concluded that the GPS must also be recording a viscoelastic response and constrained the upper mantle in this region to relatively weak viscosities of 6×10^{17}-2×10^{18} Pa s. The large difference between late Holocene RSL-derived uplift rates (3.3 to 5.1 mm/yr) and pre-2002 GPS-observed uplift rates (<0.1 mm/yr) suggests that either glacial isostatic adjustment associated with post-LGM ice loss has decreased significantly since 0.79 ka or the

RSL-derived rates reflect a localized response of the weak upper mantle to late Holocene advances and retreats, which had either decayed by 1998 or was reduced by increased accumulation following the LIA (Nield et al., 2012). This difference in pre-2002 and Holocene rates further supports a strong RSL sensitivity to late Holocene ice mass changes and a weaker relationship between late-Holocene RSL and post-LGM deglaciation.

5.3 Implications for GIA and ice-sheet models and estimates of ice-sheet mass balance changes

The relatively weak Earth structure beneath parts of the Antarctic Peninsula implied by recent GPS observations is also reflected in the rates of late Holocene RSL change observed across the Antarctic Peninsula. If this result is robust, it implies that Holocene records of RSL generated along these parts of the Antarctic Peninsula may not be as effective a tool for constraining the configuration of the LGM Antarctic ice sheet as they are across the northern hemisphere. This is important, not only because GIA model comparisons to RSL data play an important role in efforts to reconstruct past ice sheet history (e.g. Ivins and James, 2005; Whitehouse et al., 2012a; b), but also because accurate GIA predictions are therefore needed to interpret gravity-based estimates of ice loss (Shepherd et al., 2012; King et al., 2012). Our findings support earlier studies suggesting neoglacial advances can strongly influence modern rates of uplift (Ivins et al., 2000). Such effects must be included in future GIA models and accounted for in estimates of ice mass loss based on satellite gravity measurements (Ivins et al., 2011).

5. Conclusions and Future Prospects

The rate of relative sea-level fall at Torgersen Island near Palmer Station in the Antarctic Peninsula increased from 3.0±1.2 mm/yr to 5.1±1.8 mm/yr around 0.79 ka broadly contemporaneous with glacial retreat within a nearby ice piedmont. The sensitivity of these Holocene sea-levels to ice-mass changes supports GPS-based assertions of a relatively weak upper mantle beneath this parts of the Antarctic Peninsula. Furthermore, the rate of Holocene uplift at Torgersen Island, derived by correcting

RSL rates for sea-surface elevation changes and assuming negligible steric affects (5.3±1.8 mm/yr), far exceeds the pre-2002 GPS-derived uplift rates recorded at Palmer Station but is similar to the rates experienced since the breakup of the Larsen B Ice Shelf in 2002. The sensitivity of Holocene RSL rates near Palmer Station to late Holocene ice retreats suggests that these records are not recording post-LGM ice collapse but more recent ice changes, and more data are needed to constrain not only the RSL record from sites with a weak mantle in Antarctica, but also their Holocene ice history. Thus their use in constraining GIA models of the LGM ice sheet should be revisited and corrections applied to satellite data to determine ongoing mass loss should be updated accordingly.

Acknowledgements

The authors would like to thank Marissa Goerke for sharing the tide-gauge data from Palmer Station and Colin Harris for answering questions about the mapping of Torgersen Island. This project benefited from funding from the National Science Foundation Office of Polar Programs through grants OPP-0838781 and OPP-1644197 to ARS and RG. The writing of this manuscript was also made possible through the US-UK Fulbright Commission to ARS for support during his sabbatical at Durham University. This manuscript benefited by reviews from two anonymous reviewers and special issue editor Glenn Milne. The PALSEA community is also thanked for inspiring this work.

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Figure Captions:

Figure 1. Map of the northern Antarctic Peninsula illustrating the locations of places mentioned in the text.

Figure 2. A) Aerial photograph of Torgersen Island and Palmer Station Region (from GoogleEarth). B) Inset map showing the location of (A) with respect to locations shown in Figure 1 as well as Biscoe Point and Litchfield Island. C) Photograph of the raised beaches at Torgersen Island.

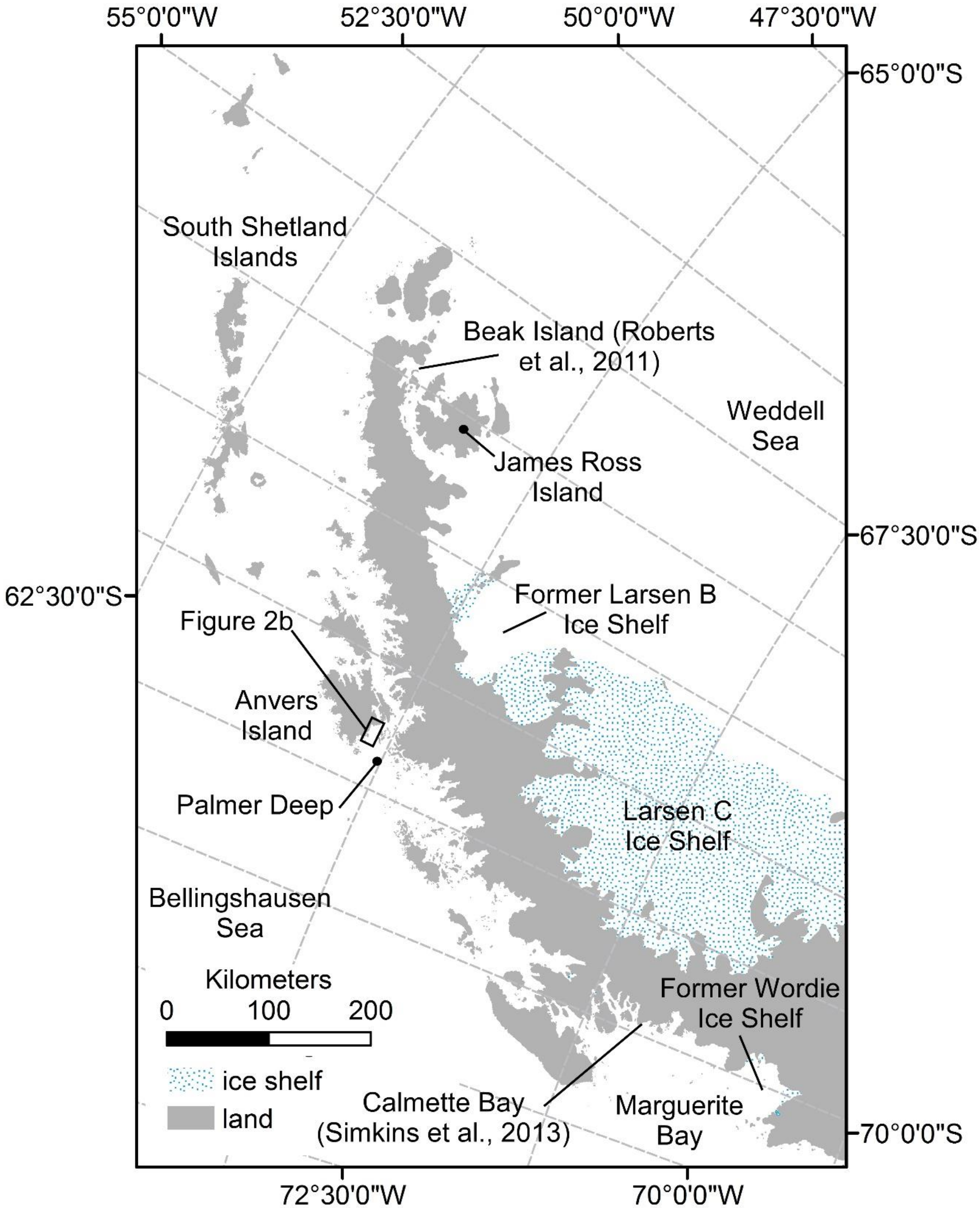
Figure 3. Model predictions of the rate of sea-surface height change (SSH) from 1.54 ka to 0.79 ka (A) and 0.79 ka to the present (B) using the ice model of Whitehouse et al. (2012b). The difference in the rate of SSH height change predicted using the ice model of Whitehouse et al. (2012) and a modified ice model with 10% less ice at 1.0 ka and 10% more ice at 0.5 ka at 1.54 ka to 0.79 ka (C) and 0.79 ka to present (D). All predictions are for Palmer Station, Antarctica.

Figure 4. Changes in uplift rates through time from GPS observations at Palmer Station (Nield et al., 2014 and Thomas et al., 2011) and Holocene relative sea-level indicators at Beak Island (Roberts et al., 2011; this study) and Togersen Island (this study). See Figure 1 for locations.

Figure 5. Glacial isostatic adjustment model results of relative sea-level changes (RSL, black lines) at Torgersen Island, deconvolved into the solid earth (DEF, red lines) and sea-surface height components (blue lines) with the original ice model of Whitehouse et al. (2012a)(A, B) and a modified ice model (C) with a 10% decrease in the ice thickness at 1.0 ka and a 10% increase in the ice at 0.5 ka (see text for details). See Figure 1 for location of Torgersen Island.

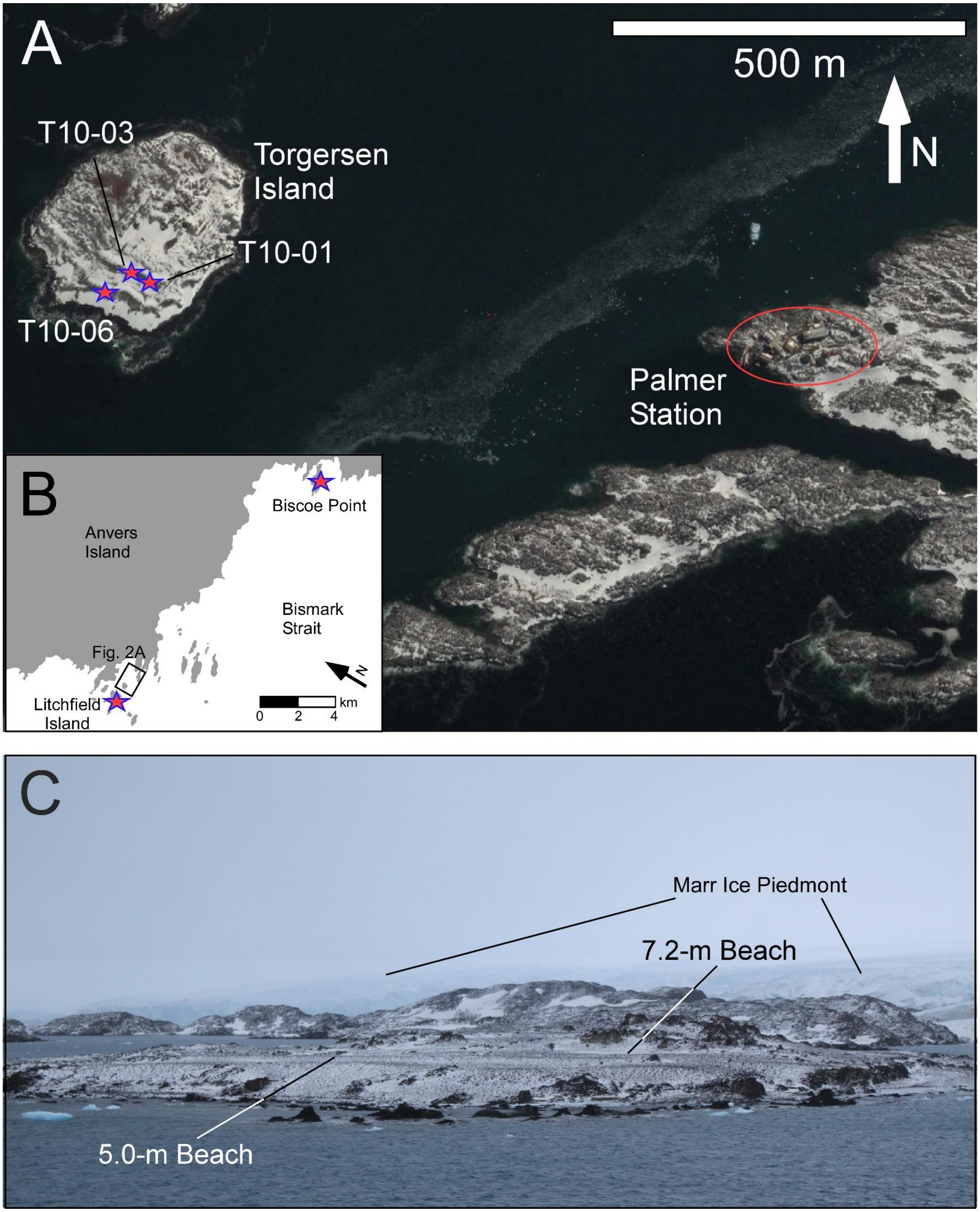
Table 1. OSL ages from the southern margin of Anvers Island, western Antarctic Peninsula

Table 2. Uplift rate calculations for Torgersen Island, Antarctic Peninsula



736

737 Figure 1.



738

739 Figure 2.

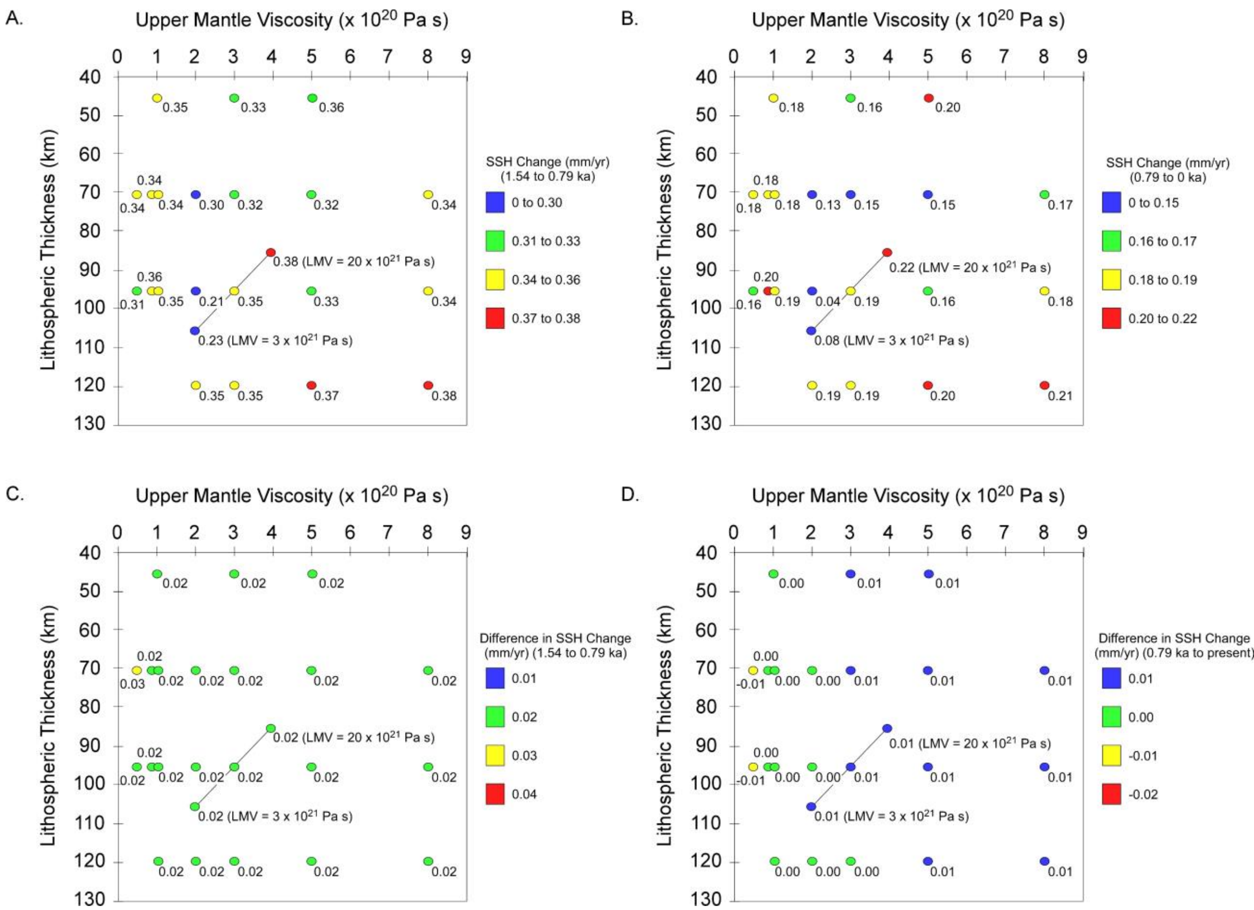
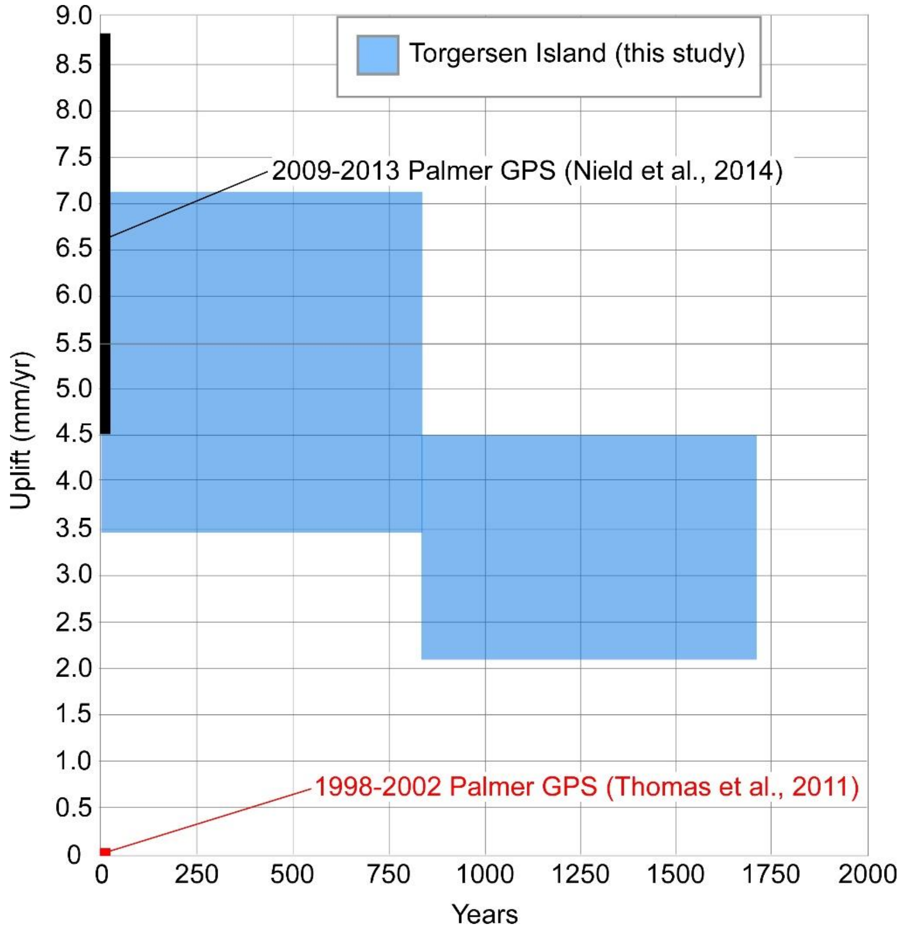

A.
Upper Mantle Viscosity (x 10^20 Pa s)
Lithospheric Thickness (km)
SSH Change (mm/yr) (1.54 to 0.79 ka)
0 to 0.30
0.31 to 0.33
0.34 to 0.36
0.37 to 0.38
0.38 (LMV = 20 x 10^21 Pa s)
0.23 (LMV = 3 x 10^21 Pa s)
B.
Upper Mantle Viscosity (x 10^20 Pa s)
Lithospheric Thickness (km)
SSH Change (mm/yr) (0.79 to 0 ka)
0 to 0.15
0.16 to 0.17
0.18 to 0.19
0.20 to 0.22
0.22 (LMV = 20 x 10^21 Pa s)
0.08 (LMV = 3 x 10^21 Pa s)
C.
Upper Mantle Viscosity (x 10^20 Pa s)
Lithospheric Thickness (km)
Difference in SSH Change (mm/yr) (1.54 to 0.79 ka)
0.01
0.02
0.03
0.04
0.02 (LMV = 20 x 10^21 Pa s)
0.02 (LMV = 3 x 10^21 Pa s)
D.
Upper Mantle Viscosity (x 10^20 Pa s)
Lithospheric Thickness (km)
Difference in SSH Change (mm/yr) (0.79 ka to present)
0.01
0.00
-0.01
-0.02
0.01 (LMV = 20 x 10^21 Pa s)
0.01 (LMV = 3 x 10^21 Pa s)

Figure 3.



Torgersen Island (this study)
2009-2013 Palmer GPS (Nield et al., 2014)
1998-2002 Palmer GPS (Thomas et al., 2011)
Uplift (mm/yr)
Years
9.0
8.5
8.0
7.5
7.0
6.5
6.0
5.5
5.0
4.5
4.0
3.5
3.0
2.5
2.0
1.5
1.0
0.5
0
0
250
500
750
1000
1250
1500
1750
2000

Figure 4.

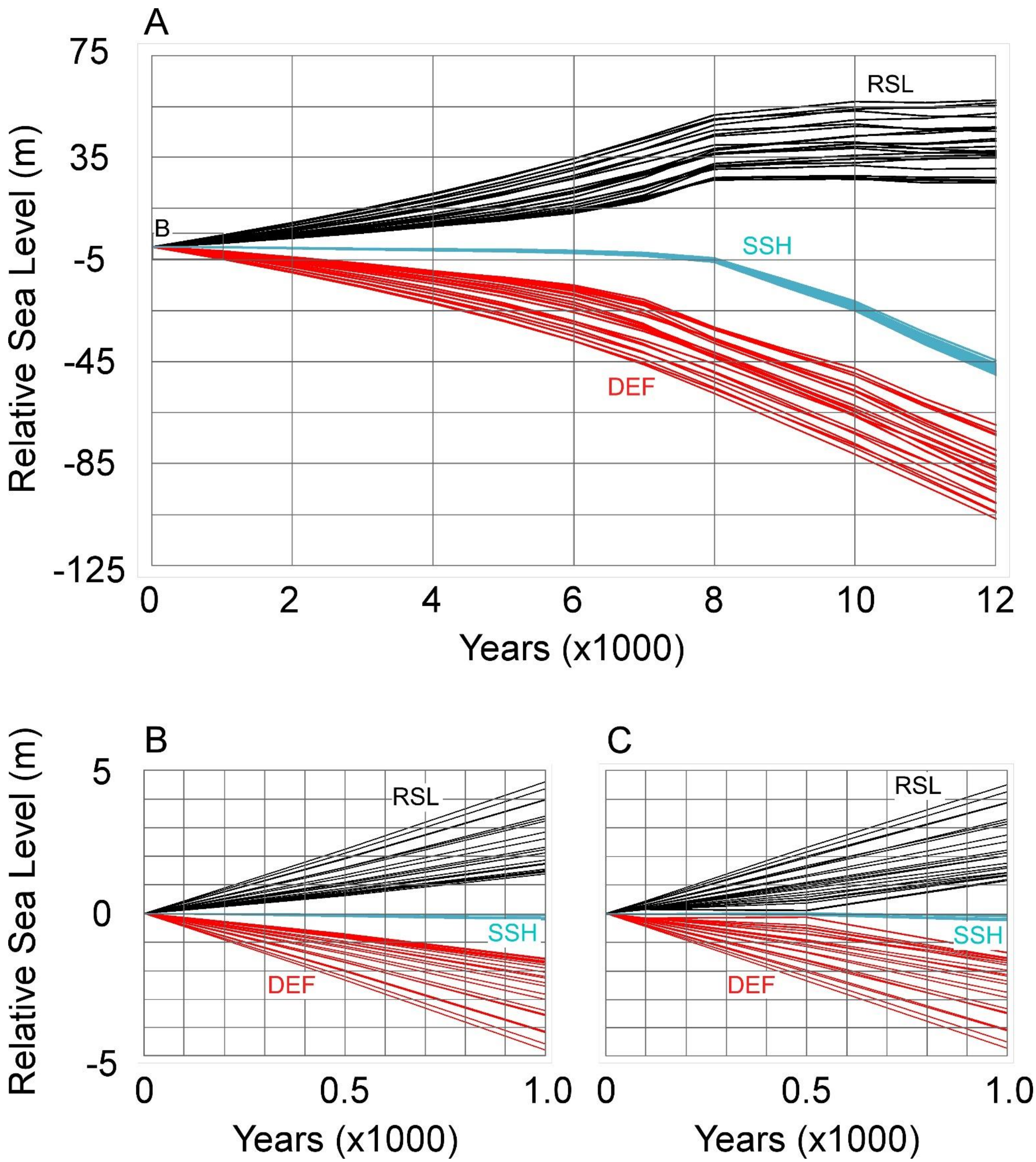

A
B
C
Relative Sea Level (m)
Years (x1000)
RSL
SSH
DEF
B
75
35
-5
-45
-85
-125
0
2
4
6
8
10
12
5
0
-5
0.5
1.0

Figure 5.

Table 1. OSL ages from Anvers Island, western Antarctic Peninsula							
Sample	**Location (Lat., Long.)***	**Elev. (m asl.)**	**Aliquotsƚ**	**D_e (Gy)**	**DR (Gy ky^{-1})**	**Age (a)‡**	**Source**
Torgersen Island							
T10-06	64.7732, 64.0765	5.0	14	1.18 ± 0.08	1.48 ± 0.14	790 ± 180	Simkins *et al.* (2015)
T10-03	64.7732, 64.0757	7.2	15	3.42 ± 0.14	2.32 ± 0.13	1480 ± 200	Simkins *et al.* (2015)
T10-01	64.7731, 64.0758	7.2	8	2.43 ± 0.21	1.11 ± 0.14	2180 ± 680	Simkins *et al.* (2015)
Litchfield Island							
LI10-08	64.7720, 64.0889	0.5	6	0	Not measured	Modern	This study
Biscoe Point							
BP10-01	64.81667, 63.81667	2	6	0	Not measured	Modern	This study
BP10-02	64.81667, 63.81667	2	6	0	Not measured	Modern	This study
BP10-05	64.81667, 63.81667	2	6	0	Not measured	Modern	This study
Elev. (elevation); De (equivalent dose); DR (dose rate); ky (thousand years); a (years)							
* In decimal degrees.							
ƚ Aliquots presented pass all standard SAR procedure tests, with the exception of modern samples that only were measured in initial dose tests.							
‡ Ages are rounded to the nearest ten years. Ages date beach formation and errors are reported as 2σ.							

Table 1.

Table 2. Uplift rate calculations for Torgersen Island, Antartic Peninsula						
Time Period (cal BP)	**RSL Rate (mm/yr)**	**Error (mm/yr)**	**SH-Correctio (mm/yr)**	**SH-Correction Erro (mm/yr)**	**Uplift Rate (mm/yr)**	**Error (mm/yr)**
0-790	5.1	1.8	0.2	0.1	5.3	1.8
790-1540	3	1.2	0.3	0.1	3.3	1.2
Difference (mm/yr)	2.1	2.2			2.0	2.2
SSH (sea surface height)						
*Error calculated as to incorporate the age uncertainty of the marine-lacustrine sedimentary contact						

Table 2.