Southern Ocean seasonal temperature and Subtropical Front movement on the South Tasman Rise in the late Quaternary


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The Subtropical Front (STF) marking the northern boundary of the Southern Ocean has a steep gradient in sea surface temperature (SST) of approximately 4°C over 0.5° of latitude. Presently, in the region south of Tasmania, the STF lies nominally at 47°C in the summer and 45°C in the winter. We present here SST reconstructions in a latitudinal transect of cores across the South Tasman Rise, southeast of Australia, during the late Quaternary. SST reconstructions are based on two paleotemperature proxies, alkenones and faunal assemblages, which are used to assess past changes in SST in spring and summer. The north-south alignment in core locations allows reconstruction of movement of the STF over the last 100 ka. Surface water temperatures during the last glaciation in this region were ~4°C colder than today. Additional temperature changes greater in magnitude than 4°C seen in individual cores can be attributed to changes in the water mass overlying the core site caused by the movement of the front across that location. During the penultimate interglacial, SST was ~2°C warmer and the STF was largely positioned south of 47°C. Movement of the STF to the north occurred during cool climate periods such as the last marine isotope stages 3 and 4. In the last glaciation, the front was at its farthest north position, becoming pinned against the Tasmanian landmass. It moved south by 4° latitude to 47°S in summer during the deglaciation but remained north of 45°S in spring throughout the early deglaciation. After 11 ka B.P. inferred invigoration of the East Australia Current appears to have pushed the STF seasonally south of the East Tasman Plateau, until after 6 ka B.P. when it achieved its present configuration.


1. Introduction

The Subtropical Front (STF) marks the northern boundary of the Southern Ocean delineating subantarctic from subtropical waters. Today, the STF is characterized by a consistent gradient of 4°C over less than 0.5° of latitude [Rintoul et al., 1997], despite the fact that the temperatures at the front vary zonally and seasonally by about 4°C [Belkin and Gordon, 1996]. In the Australian region of the Southern Ocean, summer sea surface temperature (SST) is ~16°C in subtropical waters and 12°C in subantarctic waters. In winter the SST is 8° and 12°C, respectively [Belkin and Gordon, 1996; Rintoul et al., 1997]. Frontal positions are controlled primarily by winds [McCartney, 1977], causing the STF to migrate seasonally north to south in the sectors where it is not anchored bathymetrically [e.g., Heath, 1985; Belkin and Gordon, 1996; Rintoul et al., 1997]. Because bathymetry also influences frontal location [Heath, 1980, 1981], the modern position of the STF is pulled south from its position in both the Great Australian Bight and the Tasman Sea to lie in the saddle separating the South Tasman Rise (STR) from Tasmania [Orsi et al., 1995; Belkin and Gordon, 1996]. Here, the STF oscillates between 45°S in winter and 47°S in summer [Belkin and Gordon, 1996; Rintoul et al., 1997] (Figure 1). Presently, regional SST is also influenced by the warm East Australia Current (EAC) which flows down the east coast of mainland Australia. The EAC is usually deflected to the east at about 32°S, well north of Tasmania [Ridgway and Godfrey, 1994; Ridgway et al., 2002]. However, warm rings spun off from the current occasionally travel down the east coast of the island delivering warm water as far south as the East Tasman Plateau, with recent studies suggesting that the EAC has recently strengthened [Bostock et al., 2006].

During the last glaciation, subantarctic waters of the Southern Ocean cooled by 4–9°C [Wells and Connell, 1997; Weaver et al., 1998; Nees et al., 1999; Barrows et al., 2000; Schaefer et al., 2005; Hayward et al., 2008]. In addition, Southern Ocean fronts are known to have migrated as much as 5° of latitude northward in the Indian Ocean sector [Morley and Hays, 1979; Prell et al., 1980; Morley, 1989; Howard and Prell, 1992] and 2–3° northward in the
The Subtropical Front is at 47°S, located at 47°S, 145°E; GC07 is located at 45°S, 145°14E; and the southernmost core, GC17, is located at 47°45S, 145°49E. The present location of the Subtropical Front is at 47°S in summer and 45°S in winter.

Great Australian Bight and Tasmanian regions [Wells and Okada, 1996; Passlow et al., 1997; Wells and Connell, 1997; Nees et al., 1999]. This northward frontal movement increased the surface area of the Southern Ocean during the glaciation. In contrast, where dynamic steering by bathymetry constrains frontal placement [Hayward et al., 2008], temperature gradients appear to have increased in intensity in response to the more severe climate [Weaver et al., 1998; Sikes et al., 2002].

Oceanic cooling during the last glaciation is universally accepted, but variation in SST estimates among different proxies suggest that the choice of technique affects intercomparisons as much as regional variation in climate systems. In the Australian–New Zealand region, SST reconstructions based on foraminiferal assemblages suggest that subantarctic waters cooled about 8°C [Wells and Connell, 1997; Weaver et al., 1998; Sikes et al., 2002], whereas reconstructions based on alkenones suggest cooling of about 4°–6°C [Ikehara et al., 1997; Sikes et al., 2002; Pahnke et al., 2003; Pelejero et al., 2006]. The use of multiple proxies is widely acknowledged to bring a wider understanding of paleoceanographic conditions than the use of a single technique [Mix et al., 2000; Bard, 2001; Mix et al., 2001]. To assess the proportion of variation in SST estimates attributable to different responses among techniques, this study compares estimates based on δ18O with those derived from foraminiferal assemblages. δ18O is the unsaturation index of C37 alkenones biosynthesized primarily by Emiliania huxleyi [Volkman et al., 1980]. Globally, the ratio of diunsaturated to triunsaturated compounds has been shown to change linearly with temperature [Prahll and Wakeham, 1987; Müller et al., 1998; Prahll et al., 2000]. A calibration for cold polar waters was produced in the Southern Ocean in this region [Sikes and Volkman, 1993; Sikes et al., 1997] providing a regional calibration for comparison with foraminiferal estimates. Previous comparison studies between alkenone and foraminiferal assemblage SST in other locations have been able to use the differences to assess changes in environmental parameters such as thermocline depth and seasonality that also occur with past changes in climate [Sikes and Keigwin, 1994; Chapman et al., 1996; Sikes et al., 2002].

SST reconstructions near frontal boundaries can be complicated by paleomigrations of the front. In the Southern Ocean, the temperature gradient of ~4°C across the STF is equal to the past SST change assessed in sites distal from past STF positions [e.g., Ikehara et al., 1997; Barrows et al., 2000; Sikes et al., 2002; Pahnke et al., 2003; Pelejero et al., 2006]. The similarity in magnitude makes frontal movement versus simple cooling difficult to distinguish in records from a single site. Near Australia and New Zealand at sites proximal to the STF SST is estimated to have cooled by ~6°–8°C [e.g., Wells and Connell, 1997; Weaver et al., 1998; Sikes et al., 2002; Pelejero et al., 2006] about double the temperature change estimated in distal sites. To assess what proportion of this cooling is caused by movement of the front across the site requires a latitudinal transect of sites to determine temperature gradients between sites [e.g., Howard and Prell, 1992; Sikes et al., 2002].

The focus of this study was to examine Southern Ocean last glacial to interglacial temperature change and movement of the Subtropical Front. Accordingly, three cores comprising a north-south transect underlying the modern position of the STF on the South Tasman Rise were studied. The STR is a large submarine plateau that lies south of Tasmania, Australia (Figure 1). Trending NW-SE, it is about 500 km long and lies between 1000 and 4000 m below sea level. The STR is separated from the Tasmanian landmass and the East Tasman Plateau (ETP) by a WNW trending trough more than 3000 m deep [Exon et al., 1995]. Quaternary sediments are found in localized pockets on the STR below 2000 m. Shallower than these depths the sediments are coarse foraminiferal sands winnowed by strong bottom currents of the Antarctic Circumpolar Current (ACC) which flows west to east across the rise [Connell and Sikes, 1997]. Because of these conditions, previous paleoceanographic studies on the South Tasman Rise were limited because of noncontinuous sedimentation and disturbed sections in the available deep-sea sediment cores [e.g., Wells and Connell, 1997]. A number of carefully

Figure 1. Map of the South Tasman Rise study area. Core locations are indicated by triangles. From north to south along the transect: core GC31 is located at 44°09S, 149°03E; GC07 is located at 45°09S, 146°17E; GC14 is located at 46°40S, 145°14E; and the southernmost core, GC17, is located at 47°45S, 145°49E. The present location of the Subtropical Front is at 47°S in summer and 45°S in winter.
targeted gravity cores based on detailed acoustic facies maps were collected by the RV *Rig Seismic* in 1995 (cruse RS147) [Exon *et al.*, 1995]. Cores with continuous sedimentation were chosen for this study.

### 2. Methods

[7] Three cores along an approximately north-south transect were chosen from those collected on the RV *Rig Seismic* voyage 147 [Exon *et al.*, 1995] in 1995 (Figure 1). The northern core RS147-GC07 (45° 09' S 146° 17' E 3300 m) sits at the latitude of the winter position of the STF today. The middle core RS147-GC14 (46° 26' S 145° 14' E, 3360 m) sits slightly north of the summer position of the front and the southern core RS147-GC17 (47° 45' S 145° 49' E 3300 m) sits south of the summer position of the STF today. These cores are well situated to assess north-south movement of the STF in the past. In addition to the north-south transect, a fourth core GC31 (44° 32' S 149° 03' E 3405 m) sits east of the northern core on the East Tasman Plateau. This site was chosen to assess the changing strength and influence of the EAC warm core eddies on past SST.

#### 2.1. Sampling and Sample Treatment

[8] Cores RS147-GC07, RS147-GC14, RS147-GC17, and RS147-GC31 (referred to as cores GCx hereafter) were sampled every 10 cm for CaCO₃, isotopic and alkenone analyses [Connell and Sikes, 1997]. In addition, GC07 was sampled every 2 cm from 0 to 20 cm and 90 to 130 cm, and at 1 cm intervals between 20 and 90 cm for faunal and isotopic analyses. Raw sediment samples for isotopic and faunal analyses were oven dried at 40°C, weighed and disaggregated in distilled water before wet sieving through a 63 μm mesh. Calcium carbonate content of these cores was previously reported: they are composed predominantly of calcareous foraminiferal-nanno oozes with a CaCO₃ content of 70–100% [Connell and Sikes, 1997]. All numerical data for this work are available from the World Data Center for Paleoclimatology, 325 Broadway, Boulder, Colorado; http://www.ngdc.noaa.gov/paleo/paleo.html.

#### 2.2. The δ¹⁸O, δ¹³C, and Δ¹⁴C Isotope Analyses

[9] Stable isotope analyses were performed on ~20 planktonic *Globorotalia bulloides* (300–355 μm) and 3–6 benthic *Cibicidoides* sp. (>150 μm). Prior to analysis, foraminifera were crushed and sonicated in methanol and roasted under vacuum for 1 h at 370°C to remove any organic matter. Benthic foraminiferal isotopic ratios were measured in the Geology Department stable isotope laboratory at the University of California Davis and planktonic foraminifera at the Central Science Laboratory, University of Tasmania as described by Samson *et al.* [2005]. The mean half range for replicate samples is ±0.12‰ and reported as per mil (‰) deviations from the Pee Dee belemnite (PDB) standard [Samson, 1998; Samson *et al.*, 2005].

[10] Accelerator Mass Spectrometry (AMS) radiocarbon dates were obtained from ~1 mg monospecific samples of the planktonic foraminifer *Globorotalia inflata*. In core GC07, samples were chosen from levels of peak abundances and specific faunal or isotopic changes. Dates for cores GC14, and GC17 were produced at Rafter Radiocarbon Laboratory, Institute of Geological and Nuclear Sciences, New Zealand and previously reported [Connell and Sikes, 1997]. Dates produced at Australian Nuclear Sciences and Technology Organisation (ANSTO), Australia [Samson, 1998] are reported here (Table 1). Conventional radiocarbon ages were converted to calendar ages using the marine CALIB 5.0 program (M. Stuiver et al., 2005; available at http://calib.qub.ac.uk/calib/). Radiocarbon ages older than 20 ka were calibrated to calendar ages using an oceanic ¹⁴C reservoir correction of 350 years [Druffel and Griffin, 1999] and a polynomial fit [Bard *et al.*, 1998].

#### 2.3. Alkenone Temperature Estimation

[11] Samples for $U_{37}^{KC}$ SST analyses were sampled aboard ship within 24 h of coring and stored frozen in solvent-cleansed jars at −20°C until extraction. Sediments were freeze-dried and ground using a mortar and pestle and the lipid fraction was subsequently extracted using an automatic solvent extraction device ASE-200 (Dionex) using MeOH and hexane in a 3:1 ratio at 150°C and a pressure of 2000 psi. After extraction, sediment extracts from the Holocene and marine isotope stage (MIS) 2 levels were saponified in 5% KOH in methanol (MeOH:H₂O 80:20) to obtain the neutral fraction and analyzed using capillary gas chromatography (GC) following procedures previously reported [Sikes and Volkman, 1993; Sikes *et al.*, 2005], using an HP-1 methyl silicone fused column (50 m × 0.32 mm id.) in an HP 5890 GC with a cooled on-column injector and hydrogen as the carrier gas. The temperature program was 50°C–150°C at

### Table 1. AMS¹⁴C for Planktonic Species *Globorotalia inflata* in Core RS147-GC07

<table>
<thead>
<tr>
<th>Sample Depth (cm)</th>
<th>Size Fraction (μm)</th>
<th>¹⁴C Age (years B.P.)</th>
<th>Calendar Age (years B.P.)</th>
<th>Error (1σ)</th>
<th>AMS Lab</th>
</tr>
</thead>
<tbody>
<tr>
<td>0–3</td>
<td>&gt;250</td>
<td>1,167</td>
<td>716</td>
<td>73</td>
<td>NZA 6684</td>
</tr>
<tr>
<td>40–41</td>
<td>300–355</td>
<td>9,530</td>
<td>10,870</td>
<td>120</td>
<td>OZC897</td>
</tr>
<tr>
<td>44–45</td>
<td>&gt;300</td>
<td>10,100</td>
<td>11,675</td>
<td>160</td>
<td>OZD257</td>
</tr>
<tr>
<td>48–49</td>
<td>300–355</td>
<td>11,870</td>
<td>13,304</td>
<td>180</td>
<td>OZC610</td>
</tr>
<tr>
<td>52–53</td>
<td>&gt;300</td>
<td>11,850</td>
<td>13,856</td>
<td>170</td>
<td>OZD258</td>
</tr>
<tr>
<td>57–58</td>
<td>300–355</td>
<td>11,900</td>
<td>11,420</td>
<td>110</td>
<td>OZC609</td>
</tr>
<tr>
<td>58–59</td>
<td>&gt;300</td>
<td>11,850</td>
<td>11,370</td>
<td>560</td>
<td>OZD259</td>
</tr>
<tr>
<td>65–66</td>
<td>300–355</td>
<td>13,650</td>
<td>16,242</td>
<td>140</td>
<td>OZC988</td>
</tr>
<tr>
<td>74–75</td>
<td>&gt;300</td>
<td>14,070</td>
<td>16,778</td>
<td>140</td>
<td>OZC895</td>
</tr>
<tr>
<td>80–83</td>
<td>&gt;355</td>
<td>15,360</td>
<td>18,728</td>
<td>130</td>
<td>NZA 6639</td>
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<td>89–90</td>
<td>&gt;355</td>
<td>16,450</td>
<td>19,639</td>
<td>130</td>
<td>OZC990</td>
</tr>
<tr>
<td>98.5–99.5</td>
<td>&gt;355</td>
<td>18,150</td>
<td>21,625</td>
<td>180</td>
<td>OZC991</td>
</tr>
<tr>
<td>108–109</td>
<td>&gt;355</td>
<td>23,300</td>
<td>27,520</td>
<td>340</td>
<td>OZD260</td>
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<tr>
<td>118–119</td>
<td>&gt;355</td>
<td>30,700</td>
<td>35,000</td>
<td>500</td>
<td>OZC992</td>
</tr>
</tbody>
</table>

*Sample numbers.*
30°C min⁻¹ and 150°–325°C at 3°C min⁻¹ which ensured good separation of all major constituents. Compounds were detected with a flame ionization detector (FID) and peak areas were calculated using DAPA acquisition and processing software [Sikes et al., 2002]. Extracts from MIS 3 and older were run identically to the Holocene samples by capillary GC: injected on-column using hydrogen as a carrier gas and the same temperature program, flow settings, and FID, but were analyzed using a Shimadzu GC-17A fitted with a 30 m JW Scientific DB-5 (0.32mm ID) column. Chromatograms were integrated using Shimadzu-provided software. \( U_{15} \) values were converted to SST estimates using the calibration \( U_{15} = 0.0414T - 0.156 \) which has been shown to provide more accurate temperatures in cold, polar waters, and has been calibrated for this area of the Southern Ocean in particular [Sikes and Volkman, 1993; Ikehara et al., 1997; Sikes et al., 2002; Harada et al., 2003].

The choice of temperature calibration is a critical one in this location. Temperatures calculated with the polar calibration return core top temperatures that are \( \approx 1^\circ - 2^\circ \)C too warm in the northern cores. Temperatures calculated using \( U_{15} = 0.0347T + 0.039 \) [Prahl et al., 1988] brings the core top SST of the northern core closer to modern and more in line with MAT SSTs obtained from the same sediments down-core. However, for the southern cores, SSTs are \( \approx 2^\circ \)C too cool. It appears that the polar calibration of Sikes and Volkman [1993] is more accurate where subantarctic waters bathe the site and the global calibration is a better fit for subtropical waters. Because a main objective of this study is to assess frontal movement across climate changes as well as SST, we have chosen the Sikes and Volkman [1993] calibration since subantarctic water is expected to be the dominant water mass across the study site in cooler times. Slightly elevated inferred SSTs using this calibration in the northern cores for interglacial sediments are to be expected. Temperature estimates based on the different calibrations will be offset because of the difference in intercept and small difference in slope [Prahl et al., 2000]. Accordingly, SSTs calculated using the global calibration are about 1.5°C cooler at 15°–18°C and \( \sim 3^\circ \)C cooler at \( \sim 8^\circ \)C. Errors for Southern Ocean SST estimates are \( \pm 1.5^\circ \)C [Sikes et al., 1997].

2.4. Temperature Estimation From Foraminiferal Assemblages

Planktonic foraminiferal species \( \geq 150 \mu m \) were successively split until 300–600 whole planktonic foraminifera were obtained and classified following established taxonomy [Kipp, 1976; Bé, 1977; Prell, 1985] using the updated global core top database (1219 core tops) of Prell et al. [1999]. SST estimates were obtained using the modern analog technique (MAT) which matches a down-core assemblage with modern core top samples having similar faunal assemblages [Overpeck et al., 1985; Prell, 1985; Anderson et al., 1989; Howard and Prell, 1992]. The 10 best fit core tops are chosen using squared chord distance permitting an estimate of the fit using dissimilarity coefficients (squared chord similarity). The weighted average of their associated temperatures is the temperature estimate. The analog samples also provide geographic information which can be used to interpret the past environment of the subject sample [Prell, 1985]. Dissimilarity coefficients (squared chord distance in this case) of zero are considered a perfect match, a value of 2 completely dissimilar, and dissimilarity coefficients of less than 0.2 are considered reliable. For GC07 samples, the dissimilarity of all samples was less than 0.15 [Samson, 1998]. We note here, however, that the standard deviations associated with the GC07 temperature estimates are larger than the \( 1^\circ - 1.5^\circ \)C generally associated with modern analog estimates. In GC07 the errors range from 2.0°–4.0°C for warm-season estimates in the deglaciation [Samson, 1998]. This is ascribed to the fact that GC07 lies in a zone of frontal movement where SST changes rapidly over small distances. Consequently, interseasonal movements of the frontal zone may result in the deposition of different foraminiferal assemblages over the time span of a single sample causing larger than normal standard deviations of the top 10 analogs. The modern temperatures for the database were from the International Comprehensive Ocean-Atmosphere Data Set (2008; available at http://icoads.noaa.gov/SST.html).

3. Results

3.1. Stratigraphy and Chronology

Estimates of age down-core are based on a combination of percent carbonate, \( \delta^{18}O \), and \( ^{14}C \) AMS dates when available. Below the limits of \( ^{14}C \) dating, stratigraphy was determined by tuning of \( \delta^{18}O \), and CaCO₃ levels to the SPECMAP and Marine Isotope Stage (MIS) chronology [Martinson et al., 1987; Shackleton et al., 2000; Lisiecki and Raymo, 2005]. Sedimentary carbonate content in this location shows the pattern typical of the Southern Ocean region with lower contents during glaciations [Howard and Prell, 1994; Wright et al., 1995; Carter et al., 2000; Sikes et al., 2002] (Figure 2). The changes in carbonate content and \( \delta^{18}O \) are essentially synchronous in all cores at marine isotopic stage 2.0 (14ka) [Lisiecki and Raymo, 2005]. Isotope stages 3 (24 ka), 4 (59 ka), and 5 (130 ka) [Martinson et al., 1987] are less definite than isotope stage 2 and are assigned on the basis of the best fit between variations in carbonate percentage and \( \delta^{18}O \).

The chronology for the upper portion of GC07 is based on fourteen \( ^{14}C \) AMS dates (Table 1 and Figure 2). The AMS \( ^{14}C \) ages at 48.5 cm and 52.5 cm are contemporaneous within errors, as are those at 57.5 cm and 58.5 cm. For the age model, the ages at 48.5 cm and 52.5 cm were averaged and the mean of 10.75 ka B.P. was assigned to the midpoint of the dated interval (i.e., 50.5 cm) before conversion to calendar years. As the error associated with the age at 58.5 cm was 560 years compared to 110 years for the age at 57.5 cm, the AMS \( ^{14}C \) age at 58.5 cm was not used for the age model. All numerical data presented in this paper are available as auxiliary material.

3.2. Alkenone SST Estimates

Down-core alkenone-based \( U_{25}^{K} \) SST reconstructions were generated to MIS 2 in cores GC07, GC14, and GC17
to examine the last glacial to interglacial temperature change and frontal movement across the STR. In two cores, GC07 and GC14, the $U_{37}^K$ records were extended back to MIS 5 and MIS 6 respectively to assess conditions back to the last interglacial. Alkenone levels were too low in GC31 to derive reliable SST estimates.

Figure 2. Core stratigraphy is based on a combination of CaCO$_3$ (squares with solid line), planktonic foraminiferal $\delta^{18}O$ (diamonds with dotted line), benthic foraminiferal $\delta^{18}O$ (triangles with dashed line), and $^{14}$C dates where available (double crosses). MIS boundaries are indicated by gray-shaded areas. (a) Core GC07, (b) core GC31, (c) core GC14, (d) core GC17, and (e) sedimentation rates for all cores. GC07 has 14 AMS $^{14}$C dates, and stratigraphy is very well constrained younger than $\sim$30 ka.
[17] Alkenone-based core top and late Holocene temperatures were 16°–17°C in the northernmost core (GC07), 15.5°C in the middle core (GC14), and 13°C in the southern core (GC17). These are equivalent to summer temperatures for the southern core locations, but 2°C warmer than present in the northern core. This is as expected from the choice of calibration, but within the errors for the technique (±1.5°C). In the two southern cores, core top temperatures calculated using the global calibration would be 13°C and 10°C which is 2° and 3°C too cool, respectively.

[18] The southernmost core, GC17, had an SST minimum in the last glaciation of 11.2°C. Temperatures began warming at ~20 ka B.P. to an early Holocene high of ~15.5°C at around 14 ka B.P., after which the SST dropped to a low of 13.0°C at the core top (Figure 3). Core GC14 from the middle of the north-south transect had a coolest SST of 11.2°C at the beginning of the glaciation and remained cool throughout the glaciation (11.6°C). Temperatures began to warm earlier in this core at ~21 ka B.P. compared to the more southerly site, reaching similar temperatures (15°C) by 14 ka and continued warming to a maximum temperature of 15.4°C at the core top.

[19] As with the other cores in the transect, the northernmost core GC07, had coldest temperatures (12.8°–13.0°C) during the last glaciation. An initial warming at ~21 ka B.P. was reversed until 17 ka B.P., followed by sustained warming from 17 to 13 ka B.P., at which time the $U_{37}^{K}$ temperatures in all three cores were within 1°C of each other. The early Holocene maximum in GC07 of 17.3°C, was followed by a slight SST increase to the core top high of 17.9°C.

[20] Both cores GC07 and GC14 had warmest SSTs in MIS 5 that were 2°–3°C warmer than the Holocene (ranging from ~19° to 20°C and 17.4° to 18.4°C, respectively) (Figure 3). In MIS 4, GC07 had minimum temperatures of 13.2°–13.5°C, whereas during MIS 3 this core exhibited temperatures of 16°–18°C, almost as warm as MIS 5. SSTs were unavailable in GC14 for MIS 4 because of low levels of alkenones. Coolest SST recorded in GC14 (11.2°C) were initiated in late MIS 3 and remained cool into the glaciation (11.6°C).

3.3. Foraminiferal SST Estimates

[21] Both MAT core top and late Holocene SST were ~13°C in the two northernmost cores (GC07 and GC31) (Figure 4). At 46°S (core GC14), core top SST was 12.2°C and in the southern core at 47°S (GC17) about 11.0°C. These are about 1°–2°C colder than present summer temperatures for these core locations and the same as late spring SST. Late Holocene values for GC07 were 13.3°C which is about 1°C cooler than modern summer SSTs at this site (International Comprehensive Ocean-Atmosphere Data Set).

[22] MAT-based SST reconstructions were generated across the deglaciation only to MIS 2 for the purpose of comparing these SST to $U_{37}^{K}$ in the same cores. Warm season SST reconstructions are presented here because alkenone temperatures in the Southern Ocean return a summer season SST [Sikes et al., 1997; Pichon et al., 1998; Sikes et al., 2005]. Down-core summer and winter MAT SST reconstructions are almost identical in temporal pattern but are offset by ~4°C [Samson, 1998].

[23] Faunal SST reconstructions for the southernmost core, GC17, were limited to the deglaciation and early Holocene (Figure 4). The glacial minimum SST is uncertain, but temperatures were increasing between 19 and 18 ka B.P. and reached an early Holocene maximum of ~13°C at 9 ka B.P. The record from the middle core GC14 is more complete. There is an obvious glacial minimum of 9.2°C at 22–19 ka B.P. After 19 ka B.P. temperatures rise by 4°C.
to an early Holocene maximum of \(13^\circ\text{C}\) at 12 ka B.P. (Figure 4).

[23] In the northern core of the transect, GC07, the glacial temperature minimum of \(10.1^\circ\text{C}\) occurred at 20.1 ka B.P. SST in this northern site rose earlier than at the more southerly site, beginning to increase at \(~19.5\) ka B.P. and increasing to a maximum of \(17.4^\circ\text{C}\) at \(~11\) ka B.P. (Figure 4). The high-resolution record and well constrained stratigraphy of GC07 [Samson, 1998] provides detail about the structure of the deglaciation. SSTs at the last glaciation were \(~3.5^\circ\text{C}\) colder than modern. Temperatures steadily increased from \(10.8^\circ\text{C}\) to \(15.5^\circ\text{C}\) between 19 and 16 ka B.P., but then cooled by \(~1^\circ\text{C}\) between 16 ka B.P. and 14 ka B.P., coincident with the Antarctic Cold Reversal (ACR) [Jouzel et al., 1995] before increasing again to reach maximum temperatures of \(16^\circ\text{C}–17^\circ\text{C}\) between \(~13.4\) ka B.P. and 11 ka B.P. After this, the SST drops abruptly to \(13^\circ\text{C}\) and remains around that temperature for the remainder of the Holocene.

[25] GC31 sits to the east of GC07 and was included in this study to assess the influence of the East Australia Current in the region. In this northeastern core, coldest temperatures \(~11.3^\circ\text{C}\) occurred in the glaciation and are similar to the other three cores in this study (Figure 4).

Temperatures began to rise around 20 ka B.P. warming by \(6^\circ\text{C}\), a similar magnitude increase as in GC07 and ~2°C more than the warming in GC14, or the warming suggested from the limited record in the southernmost core. After reaching an SST maximum of \(17.4^\circ\text{C}\) at 11 ka B.P., temperatures in GC31 dip to \(13.4^\circ\text{C}\) at 4 ka B.P., a pattern similar to the SST drop in GC07, but delayed in timing by nearly 6 ka (Figure 4). Overall, our glacial SST estimates agree with other faunally based SSTs from the STR [Wells and Connell, 1997; Barrows et al., 2000].

4. Discussion

4.1. Evaluation of SST Estimates

[26] It has been acknowledged that differences among independent proxies can be ascribed to nonthermal influences rather than to "problems" with an individual method and this can be qualitatively incorporated into climate interpretations to bring a wider understanding of paleoclimatographic conditions [Mix et al., 2000; Bard, 2001; Mix et al., 2001]. Nonetheless, the comparison of multiple SST records from a single core to assess a technique’s response has been rare. When compared, SST proxies tend to agree in modern conditions, because these are the conditions under which they are calibrated [e.g., Sikes and Keigwin, 1994; Chapman et al., 1996; Sikes and Keigwin, 1996; Sikes et al., 2002]. In down-core applications, SST reconstructions tend to diverge when environmental conditions other than temperature variations occurred. Overall, various proxies give similar results in the broad record, but varying results in fine detail [e.g., Sikes and Keigwin, 1994; Chapman et al., 1996; Sikes and Keigwin, 1996; Sikes et al., 2002; Barrows et al., 2007].

[27] Growth seasonality is an important nonthermal factor affecting foraminiferal assemblage SST reconstructions in the Southern Ocean. The MAT is calibrated to return an annual warm and cold season temperature even though the vast majority of the export flux for an entire assemblage occurs in a single month in the early spring associated with the spring bloom [King and Howard, 2001, 2003]. This affects both biotic and isotopic signatures of temperature [King and Howard, 2005]. In the Australian region, sediment trap studies have demonstrated that across an annual cycle, alkenone maximum export production is offset relative to foraminifera, with the maximum export of alkenones occurring in late spring to summer with some additional and significant alkenone flux occurring in late summer to autumn [King and Howard, 2001, 2003; Sikes et al., 2005]. The difference in timing between the growth seasons seen in the modern ocean suggests that changes in the relative seasonalities of the foraminifera or the alkenone-producing coccolithophorids may bias past SST estimates. Because past changes in either organisms' seasonality or response to environmental factors are difficult to reconstruct in sediment records, this bias may be noticeable only when independent paleotemperature estimates are compared [Sikes and Keigwin, 1994; Chapman et al., 1996; Sikes et al., 2002].

[28] Nonthermal factors known to affect alkenone unsaturation and \(U^{13}\text{C}_3\) accuracy are nutrient and light levels.
Glaciation indicate a northerly STF position. Convergence of the SST reconstructions in the MAT, suggesting that the front sat farther south in summer than the foraminiferal SSTs. In fact, the opposite is generally the case (Figure 5). The core top difference between the $U_{S7}^{C}$ and MAT temperature reconstructions are within $\sim 2^\circ$--$4^\circ$C of another, with alkenone reconstructions always warmer. MAT SST estimates tend to be slightly cooler than warm season today and $U_{S7}^{C}$ SST slightly warmer. This is attributed to differences in seasonality between the two organisms as evidenced from sediment trap studies and we suggest that alkenones are tracing a more fully summer season whereas the foraminifera provide a record indicative of spring conditions. Alternately, reworking of fines or light limitation may also be affecting $U_{S7}^{C}$ estimates and this cannot be fully dismissed.

Figure 5. Comparison of different proxy SST reconstructions for the last 50 ka B.P. (a) Comparison of SST reconstructions for the northern cores. GC07: $U_{S7}^{C}$ (squares with black solid line) and modern analog technique (MAT) (triangles and gray solid line). GC31 has only MAT estimates available (circles with dashed line). (b) Comparison of SST reconstructions for the southern cores. GC14: $U_{S7}^{C}$ (squares with black solid line) and MAT (circles with black dotted line). GC17: $U_{S7}^{C}$ (diamonds with gray solid line) and MAT (triangles with gray dotted line). Note that in the northern cores MIS 3 $U_{S7}^{C}$ SSTs are much warmer than MAT, suggesting that the front sit farther south in summer at that time. Convergence of the SST reconstructions in the glaciation indicate a northerly STF position.

Nutrient stress can cause increased unsaturation resulting in a cooler temperature estimate and light stress can cause an elevated temperature estimate [Prahl et al., 2003; Sikes et al., 2005]. Nutrient or light limitation can occur when waters are stratified and phytoplankton productivity shifts deeper to the nutricline. Here temperature or nutrient levels different from the surface can affect the unsaturation levels of the alkenones [Popp et al., 2006]. Another factor may be lateral transport. Alkenones are associated with fine sediment which is more readily redistributed by benthic currents than the coarse fraction which includes the foraminifera. In some locations, this has resulted in alkenones in a given samples being much older and from a remote location relative to the foraminifera in the sediment [Ohkouchi et al., 2002]. In a current-swept environment such as the STR, this effect on $U_{S7}^{C}$ has to be considered. Surface and deep currents here are in the same direction, predominantly westward flowing, which would keep any sediment redistribution generally within isotherms and unlikely to affect SST accuracy [Benthien and Müller, 2000; Sikes et al., 2002]. The exception is in core GC31, which is intermittently influenced by warm eddies spinning off from the EAC delivering waters of much warmer temperatures from the north. Currents at this core location may plausibly affect alkenone levels in the sediments [Benthien and Müller, 2000].

Temperature reconstructions in our STR cores provide evidence that local redistribution of older material does not affect the $U_{S7}^{C}$ reconstructions. In the shift from cold glacial to warmer Holocene SSTs, redistribution of older material should logically cause alkenone temperatures to remain colder longer and $U_{S7}^{C}$ SSTs to warm later than foraminiferal SSTs. In fact, the opposite is generally the case (Figure 5). The core top difference between the $U_{S7}^{C}$ and MAT temperature reconstructions are within $\sim 2^\circ$--$4^\circ$C of another, with alkenone reconstructions always warmer. MAT SST estimates tend to be slightly cooler than warm season today and $U_{S7}^{C}$ SST slightly warmer. This is attributed to differences in seasonality between the two organisms as evidenced from sediment trap studies and we suggest that alkenones are tracing a more fully summer season whereas the foraminifera provide a record indicative of spring conditions. Alternately, reworking of fines or light limitation may also be affecting $U_{S7}^{C}$ estimates and this cannot be fully dismissed.

Accepting the interpretation that $U_{S7}^{C}$ SST estimates produce a summer SST and that the warm season MAT may record spring SST in the past, informs the choice of the polar $U_{S7}^{C}$ calibration in this study. The polar $U_{S7}^{C}$ calibration results in SST reconstructions that are warmer than MAT SST, a result that is considered more realistic given the seasonality of the organisms. We note here that the temperature estimates based on the different $U_{S7}^{C}$ calibrations are subparallel such that the differences are greater at cooler temperatures. SSTs calculated using the global calibration [Prahl et al., 1988] are cooler by $1.5^\circ$C at $15^\circ$--$18^\circ$C and $3^\circ$C cooler (about twice as much) at the coldest temperatures in this study ($\sim 8^\circ$C). Use of the global calibration [Prahl et al., 1988; Müller et al., 1998] results in SST estimates that are colder than MAT SST at all times and up to $8^\circ$C colder during the deglaciation, which we do not consider realistic. An alternate explanation for “cool” alkenone SSTs could be nutrient limitation of alkenones throughout the glaciation and deglaciation. However, evidence from benthic faunal abundances suggests that productivity was enhanced on the South Tasman Rise during this time [Nees et al., 1999].

Our $U_{S7}^{C}$ values agree well with two other alkenone-based SST records from the South Tasman Rise area that sit very near cores GC17 [Ikehara et al., 1997] and GC31 [Pelejero and Grimalt, 1997]. Both of these studies used the global calibration. The absolute temperatures in those studies agree with $U_{S7}^{C}$ SST calculated in our cores using those calibrations. Temperature minimums in the glaciation were $\sim 7^\circ$C and southern site and $10^\circ$C for the northern site. Regardless of the calibration, these studies and ours agree...
that the total magnitude of warming in this region was \(\sim 4^\circ - 5^\circ\)C [Ikehara et al., 1997; Pelejero et al., 2006].

4.2. Past Positions of the Subtropical Convergence

[32] The paleolocation of a front is accurately determined by establishing the temperature gradient across a latitudinal transect [Howard and Prell, 1992; Weaver et al., 1998; Sikes et al., 2002]. This does not require the assumption that either the temperature gradient or that temperatures north and south of the front were the same in the past [Wells and Okada, 1996; Ikehara et al., 1997; Passlow et al., 1997; Wells and Connell, 1997]. Following this approach, an increase in the temperature differences between the cores suggests frontal placement between them whereas coalescing of temperatures implies the core sites were within the same water mass.

[33] In the last glaciation, all STR cores had a narrow SST range of \(\sim 3^\circ\)C and temperatures from both SST methods overlapped (11\(^\circ\) - 14\(^\circ\)C for \(U_{27}^S\) and 9\(^\circ\) - 11\(^\circ\)C for MAT (Figures 3 and 4)). This gradient is narrower than the modern temperature range and these SSTs are the minimum seen in all cores. This suggests that the STF sat north of all sites and that the spring-summer seasonal SST difference was smaller during the glaciation (Figure 6c).

[34] After the synchronous temperature minimum in the early glaciation, the timing and specifics of the two proxies diverge in detail (Figure 5). We interpret the differences between the MAT and \(U_{27}^S\) proxies as delineating the frontal positions in two seasons across the deglacial transition in climate. At the two southern sites, GC17 and GC14 (46\(^\circ\) - 47\(^\circ\)S), \(U_{27}^S\) SST warming began around 20 - 21 ka B.P., about 3 ka earlier than MAT SST, with this temporal offset greatest in GC14 (Figure 5b). This warming was unaccompanied by a change in the temperature gradient between cores and has been observed elsewhere on the southern STR [Ikehara et al., 1997]. This was followed \(\sim 2\) ka later by a synchronous rise in MAT SST in the southern cores. The consistent gradient among cores suggests the front remained well north on the STR whereas the timing lag between techniques indicates summers warmed early and spring remained cool at the beginning of the transition, possibly because winters remained more severe [Samson, 1998].

[35] A very rapid and intense deglacial warming of nearly 7\(^\circ\)C in MAT SST in the northern and northeastern sites (GC07 and GC31) at 19 ka B.P. contrasts strongly with the gradual 4\(^\circ\)C increase in the two southern sites (Figure 5). The MAT warming in the northern sites led slightly, but was essentially synchronous with the southern sites (Figure 5a). Consequently, the temperature difference between the MAT SST in the northern and southern cores jumped from about 3\(^\circ\)C to about 8\(^\circ\)C during the deglaciation, indicating that the front moved south and sat between the northern and southern pairs of cores (Figure 4) remaining between 45\(^\circ\)S and 46\(^\circ\)S in spring until about 11 ka B.P. (Figure 6d). In contrast, \(U_{27}^S\) SSTs at the northern site warmed slightly at 21 ka B.P., then cooled before warming by \(- 4^\circ\)C beginning at 17 ka B.P. (Figure 5a). Delayed warming of \(U_{27}^S\) SST has been observed previously on the East Tasman Plateau [Pelejero et al., 2006]. This difference in timing and magnitude of warming between the proxies in the northern sites indicates that the although the front shifted south early in the deglaciation, summer SST in the subtropical waters that bathed the northern sites remained relatively cool until about 2 ka later. From 17 to 12 ka B.P. the alkenone SST records in all cores coalesce, placing the STF north of all sites in summer. Through the latter part of deglaciation, the front remained south of 47\(^\circ\)S in summer, at least \(\sim 2\) further south than in spring. Overall, the deglaciation is characterized by a strong pulse of regional warming evidenced by the southerly position of the front and warming of the water masses on either side (Figures 5 and 6).

[36] Around 11 ka B.P. \(U_{27}^S\) MAT SST in GC07 dropped by 4\(^\circ\)C producing temperature differences among the cores similar to the modern (Figures 4 and 5). The striking early Holocene maximum seen in northern MAT SST is not well represented in either the \(U_{27}^S\) SST or the southern MAT SST (Figure 5). This places the STF north of 46\(^\circ\)S in spring but south of all sites during summer in the early deglaciation (Figure 6d). In contrast, MAT temperatures in GC31 on the ETP remained warm until \(\sim 6\) ka B.P. The delayed cooling on the ETP is attributable to strengthening and reinitiation of the EAC around 11 ka B.P. [Bostock et al., 2006]. In addition, \(U_{27}^S\) SST in the southern most core (GC17) cooled gently throughout the Holocene. Thus, it appears that after a strong deglacial warming pulse during which the STF sat around 47\(^\circ\)S in summers while remaining north of 45\(^\circ\)S in spring, the STF moved north at the beginning of the Holocene. However, it remained restricted from the ETP by enhanced EAC eddies until 6 ka B.P. (Figure 6d).

[37] Reconstruction of the frontal position across 45\(^\circ\) - 46\(^\circ\)S during MIS 3 and 5 are possible from \(U_{27}^S\) in cores GC07 and GC14 (Figure 3). In MIS 5 SST fluctuated strongly in the 2 cores but were essentially similar, indicating the front was located predominantly south of 46\(^\circ\) - 47\(^\circ\)S during the last interglaciation (Figure 3). The latter part of MIS 3 is distinguished by a greater than 6\(^\circ\)C temperature difference between the two cores. This suggests that not only did the front sit between them (north of 46\(^\circ\)S) but also that there was an enhanced frontal gradient as the area entered the last glaciation (Figure 6b). An intermediate southerly position of the front in MIS 3 fits with a warm event identified regionally at 39 ka B.P. [Barrows et al., 2007]. The southern frontal position in MIS 5 supports numerous paleotemperature reconstructions indicating MIS

Figure 6. Reconstruction of the Subtropical Front positions for different climates in the past. (a) During MIS 5, the STF was in a position similar to modern and in the most southerly position recorded. (b) During MIS 3 the position of the front largely resided between sites GC14 and GC17, putting its mean position at around 46\(^\circ\)S in summers. (c) The last glaciation STF sat at its most northerly position in the last \(\sim 100\) ka, when it was north of all sites and pinned against the Tasmanian landmass. (d) The front reached its most southerly position early in the deglaciation when strong warming pushed its position south of 47\(^\circ\)S in summers and north of 46\(^\circ\)S in spring. After this pulse of warming, the front relaxed to the north at \(\sim 11\) ka B.P., except on the East Tasman Plateau where initiation and strengthening of the EAC kept SSTs there elevated until 6 ka B.P.
Figure 6

(a) MIS5 position

(b) MIS3 position

(c) Glacial position

(d) Early Holocene position

Most southerly position

Southerly position

Glaciar position

Early Holocene position

Most northerly position

Southerly position

Strong EAC eddies
5 in the Southern Ocean was warmer than the present [Sikes et al., 2002; Pelejero et al., 2006; Barrows et al., 2007].

[38] In summary, we have evidence for a consistent SST gradient across the Subtropical Front of ~4°C throughout the last ~100 ka. The multiproxy record suggests that south of Tasmania, the STF moved at least 1° north in the last glaciation, where it would have been trapped against the Tasmanian landmass, expanding the aerial extent of the Southern Ocean. During the deglaciation, the front moved at least 2–3° south of its present location to 47°S and perhaps as far as 48° south in summers, before relaxing back in the mid-Holocene to a mean position of 46°S in the summer and 45°S in the spring. Indications are that summers warmed earlier in the transition with springs remaining cool through the early deglaciation. Bathymetry does not appear to control or confine the movements of the STF over the STR.

4.3. Regional Climate in the Middle to High Southern Latitudes of the Australian Region

[39] The temperature minimum across the STF occurred at 20–24 ka B.P. This early glacial SST minimum has been seen in numerous subantarctic alkenone SST reconstructions across the region [Ikehara et al., 1997; Sikes et al., 2002; Pahnke and Sachs, 2006; Pelejero et al., 2006; Barrows et al., 2007]. SST in all these studies began to increase before the initiation of the deglacial isotopic shift suggesting that summers were warm at the end of the glaciation across the region. Our SST reconstructions also agree well with temperature reconstructions from several methods across the Australia–New Zealand region for MIS 5 supporting the consensus that the penultimate interglacial was 2°–4°C warmer than modern [Ikehara et al., 1997; Sikes et al., 2002; Pahnke and Sachs, 2006; Pelejero et al., 2006; Barrows et al., 2007].

[40] When temperature changes due to frontal movements across the sites are accounted for, the warming of the individual water masses either side of the STF was ~4°C between the last glaciation and the present. Our alkenone-based reconstructions agree well with both foraminiferal-based glacial SST values [Barrows et al., 2000] and US34 SST records from the area that show total warming of ~4°–5°C [Ikehara et al., 1997; Sikes et al., 2002; Pelejero et al., 2006]. In contrast, terrestrial temperature estimates from Tasmania suggest that glacial cooling was slightly greater (5°–6°C) and warming was unidirectional with no evidence for an early Holocene temperature maximum [e.g., Colhoun et al., 1994]. However, the degree to which temperature alone drives glacial advances and pollen records remains a debated point [e.g., McGlone, 1995]. Overall, warming of the water masses is in broad agreement with regional SST reconstructions when frontal proximity to core locations is accounted for [e.g., Ikehara et al., 1997; Barrows et al., 2000; Sikes et al., 2002; Pahnke and Sachs, 2006; Pelejero et al., 2006; Barrows et al., 2007].

5. Conclusions

[41] Our transect of cores from the South Tasman Rise provides a record of SST change and latitudinal movement of the Subtropical Front in the Australian sector of the Southern Ocean. Dual SST reconstructions from foraminiferal and alkenone proxies delineate past seasonal variation in SST and STF position. Regional SST in the last glaciation was ~4°C colder than today. During the penultimate interglaciation SST was ~2°C warmer and the STF was largely positioned south of 47°S. In cool climate periods such as MIS 3 and MIS 4, the STF had a relatively northerly position, around 45°–46°S. In the last glaciation, the frontal position was so far north it became pinned against the Tasmanian landmass. Throughout the early deglaciation, the STF was 2°C farther south than its glacial location in the spring but moved an additional 2° farther south in summers. Invigoration of the EAC around 11 ka B.P. pushed the boundary of the STF south of the East Tasman Plateau. Overall, SST fluctuations across the STF over the last 100 ka reflect both regional climate changes and localized movement of the Subtropical Front across the South Tasman Rise.

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