# The 100 000-Yr Cycle in Tropical SST, Greenhouse Forcing, and Climate Sensitivity

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ABSTRACT

The key scientific uncertainty in the global warming debate is the equilibrium climate sensitivity. Coupled atmosphere–ocean general circulation models predict a wide range of equilibrium climate sensitivities, with a consequently large spread of societal implications. Comparison of models with instrumental data has not been able to reduce the uncertainty in climate sensitivity. An alternative way to gauge equilibrium climate sensitivity is to use paleoclimatic data. Two recent advances, the development and application of proxy recorders of tropical sea surface temperature (SST) and the synchronization of the deep-sea and Antarctic ice-core time scales, make it possible to directly relate past changes in tropical SST to atmospheric carbon dioxide ($CO_2$) levels. The strong correspondence of a proxy SST record from the eastern equatorial Pacific and the Vostok $CO_2$ record suggests that varying atmospheric carbon dioxide is the dominant control on tropical climate on orbital time scales. This effect is especially pronounced at the 100 000-yr cycle. Calibration of the $CO_2$ influence via tropical SST variability indicates a tropical climate sensitivity of 4.4°–5.6°C (errors estimated at ± 1.0°C) for a doubling of atmospheric $CO_2$ concentration. This result suggests that the equilibrium response of tropical climate to atmospheric $CO_2$ changes is likely to be similar to the upper end of available global predictions from coupled models.

## 1. Introduction

The similarity of Antarctic proxy air temperature and atmospheric carbon dioxide ($CO_2$) records from the Vostok ice core is striking (Petit et al. 1999; Cuffey and Vimeux 2001) and suggests that $CO_2$ is playing a direct role in forcing the observed climate variations over the ice sheet. Such past relationships between temperature and greenhouse forcing can be used to determine climate sensitivity—the surface temperature response to a prescribed radiative forcing when all feedbacks, such as water vapor and clouds, are included. But there are inherent difficulties in using the Antarctic climate to gauge global climate sensitivity, because factors unique to the ice sheet, such as elevation, ice albedo, and sea ice, must play an important local role (Genthon et al. 1987; Lorius et al. 1990). Climate sensitivity can be computed from global temperature change and forcing estimates for the last glacial maximum (LGM; Hansen et al. 1984; Hoffert and Covey 1992; Hansen et al. 1998), but this approach cannot be applied to longer time series due to the lack of time-dependent estimates of global temperature change and forcing.

The tropical oceans, because they are removed from the direct climate impact of continental ice sheet build up during glaciation, provide an analog for equilibrium climate response to varying $CO_2$ (Crowley 2000). In addition, modeling results for the glacial oceans support the hypothesis that $CO_2$ variations are the dominant source of radiative forcing in tropical ocean regions, although not all of the studies explicitly separate out the effect of $CO_2$ from other factors (Broccoli and Manabe 1987; Hewitt and Mitchell 1997; Weaver et al. 1998; Broccoli 2000) [but see Ganopolski et al. (1998) for a somewhat different view].

Previous research has demonstrated that temperature changes in the tropical oceans during glaciation were significantly smaller than is observed in oceanic areas proximal to the continental ice sheets (CLIMAP 1981; Crowley 2000). In fact, changes in tropical SST during the ice ages were small enough that they largely escaped detection until the development of two sensitive geochemical temperature proxies—alkenone unsaturation ratios and Mg/Ca in planktonic foraminifera (Pelejero et al. 1999; Lea et al. 2000). Recent research based on these two independent techniques suggests that the tropical oceans cooled by 2.8° ± 0.7°C at the LGM, 21 000 yr before present (BP) (Hastings et al. 1998; Pelejero et al. 1999; Rühlemann et al. 1999; Schneider et al. 1999; Lea et al. 2000; Nürnberg et al. 2000; Kienast et al. 2001; Stott et al. 2002; Lea et al. 2003; Rosenthal et al. 2003; Visser et al. 2003). The most important terrestrial constraint on LGM tropical cooling, the lowering of tropical mountain snowlines, is consistent with a ~3°C cooling at the sea surface (Greene et al. 2002).

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An important aspect of the available tropical paleoclimatic data is that all of the longer records, regardless of method, agree in showing the dominance of the SST signal at a period of 100 000 yr. In addition, a number of the records show that the phasing of the 100-kyr SST signal leads that of the associated ice volume signal (Pisias and Mix 1997; Schneider et al. 1999; Lea et al. 2000; Nürnberg et al. 2000; Visser et al. 2003). The strength and timing of the 100-kyr-SST signal is an important clue to the source of tropical temperature variations because recent work suggests that the 100-kyr signal in late quaternary paleoclimate records comes from the influence of atmospheric carbon dioxide variations (Shackleton 2000).

In this study I use a 360-kyr-long SST record from the Cocos Ridge, in the northeastern tropical Pacific, to evaluate the relationship between greenhouse forcing and tropical temperatures. The similarity of this record to SST records from the western tropical Pacific (Lea et al. 2000; Stott et al. 2002; Rosenthal et al. 2003; Visser et al. 2003) suggests that the Cocos Ridge site records basinwide tropical Pacific SST variations. The two main objectives of this paper are 1) to test the hypothesis that atmospheric $CO_2$ variations are the dominant control on tropical SST on orbital time scale, and 2) to compute a (tropical) climate sensitivity based on the tropical SST response to known greenhouse forcing.

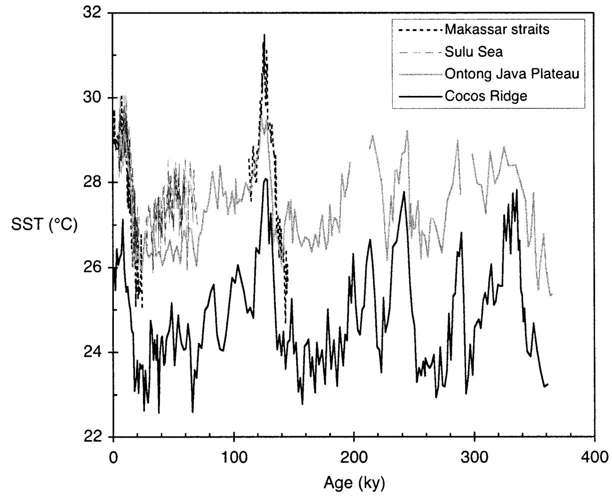


FIG. 1. Comparison of the Cocos Ridge SST record (2°15.5′N, 90°57.1′W, this study) with previously published SST records from the western tropical Pacific. The records are from the Makkassar Straits (4°41.3′S, 117°54.17′E; Visser et al. 2003), the Sulu Sea (8.8°N, 121.3°E) (Dannenmann et al. 2003; Oppo et al. 2003; Rosenthal et al. 2003), and the Ontong Java Plateau (0°19.1′N, 159°21.7′E; Lea et al. 2000). The comparison indicates that the amplitude of glacial–interglacial temperature variations is similar between sites.

## 2. The Cocos Ridge SST record

### *a. Data description and comparison to western tropical Pacific sites*

A detailed 360-kyr record of past tropical SST variations in the warm [modern SST = 26.2°C (Levitus and Boyer 1994)] eastern equatorial Pacific (EEP) waters overlying the Cocos Ridge was previously obtained using Mg/Ca determinations on the surface-dwelling foraminifera *Globigerinoides ruber* from core TR163-19 (2°15.5′N, 90°57.1′W; 2348 m; Lea et al. 2000). The magnitude of the SST variations at the Cocos Ridge site has been validated by demonstrating their compatibility with the companion oxygen isotope record, which contains a partial temperature influence, from the same samples (Lea et al. 2002). The previously published dataset has been augmented with 76 new analyses in 39 intervals, following the method described in Lea et al. (2000) but using a higher-precision Finnigan MAT Element2 inductively coupled plasma mass spectrometer (ICP-MS; Lea et al. 2003). Measured Mg/Ca values are converted to SST using the Pacific *G. ruber* equation in Dekens et al. (2002). The changes in the new merged dataset are very subtle and mostly consist of slightly reduced variability in marine isotope stage (MIS) 5 and 9.

The Cocos Ridge site lies just north of a region with large SST gradients (Levitus and Boyer 1994), and a reasonable concern would be that a gradient shift in the past due to circulation shifts might amplify the extent of tropical cooling as recorded at this site. Records from several sites in the western tropical Pacific warm pool region are now available for comparison (Fig. 1). Temperature changes over the last two glacial terminations are similar between the Cocos Ridge site and the western Pacific sites, suggesting that the Cocos Ridge SST record is representative of Pacific-wide changes.

### *b. A new time scale for the Cocos Ridge record*

The Cocos Ridge SST record bears a strong resemblance in both form and timing to the Vostok ice deuterium/hydrogen (D/H) record, which is a proxy of air temperature over Antarctica (Petit et al. 1999; Lea et al. 2000). This broad similarity suggests a common climate forcing, with the most likely candidate being atmospheric carbon dioxide, which is well mixed throughout the atmosphere and has the potential to link climate change in regions as remote as the Tropics and poles (Genthon et al. 1987; Lorius et al. 1990). The Mapping Spectral Variability in Global Climate Project (SPECMAP) time scale used previously for the Cocos Ridge SST record (Lea et al. 2000) diverges significantly from the Vostok ice core extended glacialogical time scale prior to 260 kyr BP (Petit et al. 1999). Therefore, a detailed comparison of tropical temperatures and atmospheric $CO_2$ requires a common time scale for the two archives.

Shackleton (2000) developed a common time scale for the deep-sea sedimentary record and the Vostok ice core by independently tuning an eastern equatorial Pacific benthic foraminiferal oxygen isotope record (core

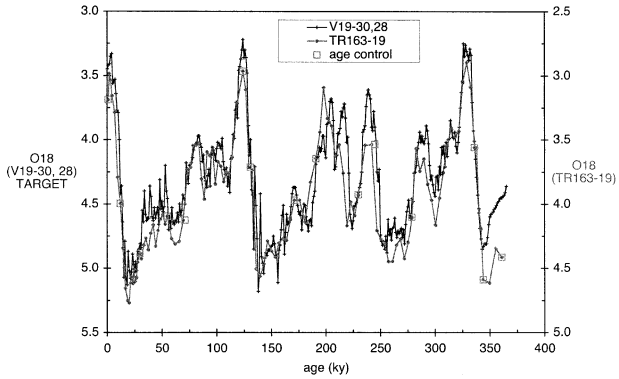


FIG. 2. Alignment of benthic oxygen isotopes (*Uvigerina* spp.) from Cocos Ridge core TR163-19 (Lea et al. 2002) to the oxygen isotope record of the same species from core V19-30 on the southeastern flank of the Carnegie Ridge (Shackleton et al. 1983). The time scale for core V19-30 is based on Shackleton et al. (2000). Alignment was done graphically by establishing tie points at well-defined transitions and clear climate events. Points used for alignment are shown by squares—a minimum number of tie points was used, resulting in a relatively constant sedimentation rate throughout the entire sequence (rms = ±25%). Shifts relative to the original chronology of TR163-19 (Lea et al. 2000) are most significant in MIS 5 and 7. Note that the benthic isotope records from these sites cannot be taken as ice volume records because ~40% of the amplitude is due to deep-sea temperature changes (Shackleton 2000; Martin et al. 2002).

V19-30; Shackleton et al. 1983) and the Vostok air bubble oxygen isotope record to orbital targets. Because benthic oxygen isotope records within a given region are expected to be quite similar, it is possible to graphically align the benthic oxygen isotope records of TR163-19 (Lea et al. 2002) and core V19-30 (Fig. 2). This alignment produces an independently aligned time scale that allows, for the first time, a direct comparison of tropical SST and atmospheric $CO_2$ variability (Fig. 3). The alignment of the benthic oxygen isotope records was achieved with a minimum number of tie points, and the resulting sedimentation rates vary by only ±25% (rms). The Cocos Ridge SST data on the new time scale are available from the author and from the NOAA World Data Center for Paleoclimatology (see information online at http://web.ngdc.noaa.goc/paleo/data.html).

### *c. Comparison to Vostok atmospheric $CO_2$ record*

The Cocos Ridge SST and atmospheric $CO_2$ records show a striking correlation throughout the 360-kyr sequence, with an overall cross correlation of $r = 0.84$ (Fig. 3). The correspondence between the two records is clear at both the longer 100-kyr periodicity, which dominates both records, and at the shorter orbital period of 41 kyr. The only time interval where the records clearly diverge is during MIS 5e and 5d, ~130–110 kyr BP, when the $CO_2$ record indicates a long plateau of high levels while the SST record indicates a clear dip

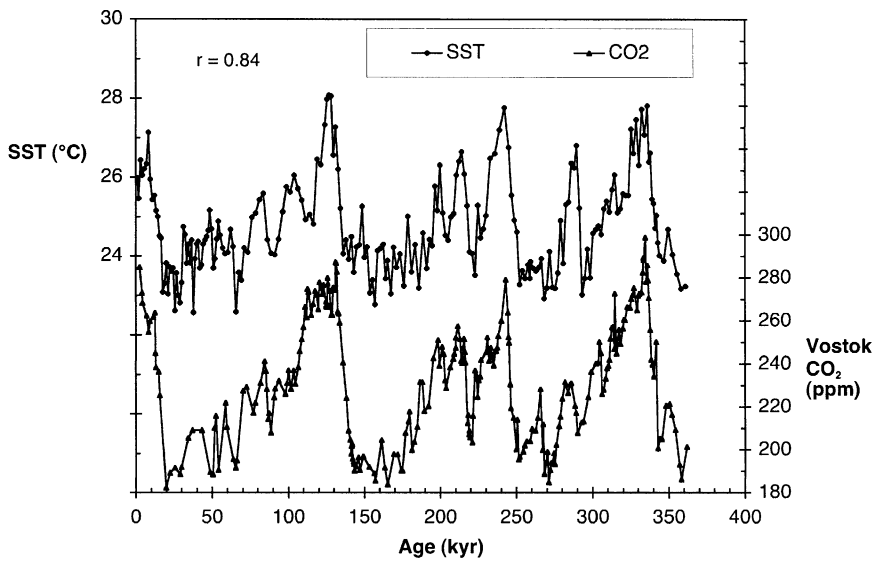


FIG. 3. Comparison of tropical Pacific sea surface temperature (SST), derived from Mg/Ca determinations on surface-dwelling planktonic foraminifera in core TR163-19 on the southwest Cocos Ridge (Lea et al. 2000), with atmospheric $CO_2$ levels determined from the Vostok ice core, east Antarctica (Petit et al. 1999). The time scale for Vostok $CO_2$ was derived from Shackleton (2000); the time scale for the Cocos Ridge core is based on aligning the benthic oxygen isotope record (Lea et al. 2002) from core TR163-19 to nearby core V19-30, which was previously aligned to Vostok (Shackleton 2000). The modern SST at the tropical Pacific site is 26.2°C (Levitus and Boyer 1994). The TR163-19 SST record is based on Mg/Ca analyses published in Lea et al. (2000), supplemented by additional analyses in ~15% of the intervals. SSTs are calculated from the *G. ruber* equation in Dekens et al. (2002).

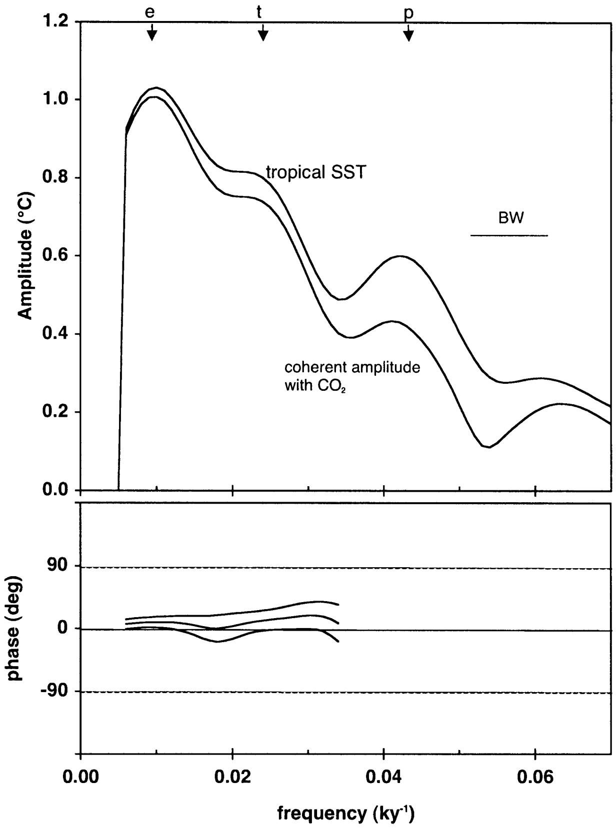


FIG. 4. Amplitude spectrum of tropical SST, showing distribution between the three major earth orbital periods: $e$ = eccentricity, $t$ = tilt, $p$ = precession. The fraction of the tropical amplitude that is coherent with Vostok $CO_2$ is shown. The coherencies between the two records at eccentricity and tilt are 0.98 and 0.94, respectively. The lower plot shows the phase relationship between the two time series, where the results indicate that the two records are in phase at eccentricity and tilt. Results based on Blackman-Tukey method (Arand package) with 1/3 lags.

to colder levels during MIS 5d. The long plateau in 5e in Vostok has previously been singled out as anomalous (Jouzel et al. 1993; Cuffey and Vimeux 2001) and appears to be shorter in a new record from the Dome Fuji Station (Watanabe et al. 2003).

Cross-spectral analysis confirms the coherence with no discernable phase lag of the 100- and 41-kyr periodicity in the two records (Fig. 4; Table 1). The amplitude spectrum of the SST and $CO_2$ records indicates that the shared coherence at 100 kyr is 98%. At this period, the peak coherent amplitudes of the SST and $CO_2$ records are 1.0°C and 28 ppmv, respectively. Comparison of the extracted deep-sea temperature component of the V19-30 benthic oxygen isotope record and the $CO_2$ record using a similar analysis yielded a coherent 100-kyr amplitude of 1°C and 34 ppmv, respectively (Shackleton 2000). The coherent 100-kyr amplitude in the Cocos Ridge SST record is likely to be a consequence of the influence of atmospheric $CO_2$ on tropical climate.

It is important to note that the amplitude of SST change at 100 kyr is not unusually high at the Cocos Ridge site. For example, a longer record from the Ocean Drilling Program (ODP) hole 806B in the western Pacific, in the center of the warm pool, has a slightly higher 100-kyr amplitude of 1.1°C (using the data and age model in Lea et al. 2000; Fig. 1). This site does have clearly reduced amplitude at the other major orbital periods, so the overall impression of the record is of a somewhat reduced amplitude relative to the Cocos Ridge site (Lea et al. 2000).

The amplitude spectrum of the Cocos Ridge SST record indicates that a significant fraction of the variability at shorter orbital periods (41 and 23 kyr) is not coherent with $CO_2$ (Fig. 4). The two records are not statistically coherent at the frequency of precession, and the tropical SST record has significantly higher variance at 23 kyr. This is also apparent in the graphical data comparison (Fig. 3); the three large oscillations of SST during MIS 5 (~130–80 kyr BP) are either reduced or absent in the $CO_2$ record. The implication is that some of the 23-kyr variability in the SST record comes about from other climate influences; one possible candidate is the influence of precession on the seasonal distribution of radiation in the Tropics, which could change the tendency of the tropical Pacific toward warm El Niño versus cold La Niña phases (Clement et al. 1999). Multivariate analysis, however, suggests that the additional influence is likely to come from the high latitudes, presumably through ice volume (see below).

It has previously been suggested that the tropical temperature changes themselves could have caused the atmospheric $CO_2$ change (Visser et al. 2003). Carbonate equilibrium calculations, box models, and ocean GCMs demonstrate that the expected atmospheric $CO_2$ response to the observed low-latitude glacial cooling is too small to explain the observed ice core $CO_2$ change (Archer et al. 2000). Therefore, warming in the Tropics in response to rising atmospheric $CO_2$ probably provided only a small feedback through the decreased solubility of $CO_2$ in warmer seawater.

TABLE 1. Spectral analysis results for comparison of the Cocos Ridge SST record (TR163-19) to the Vostok atmospheric carbon dioxide record. Confidence limits for phase lags are at the 95% level.

| | Tot amplitude | | Coherent amplitude | | | Phase lag | |
|---|---|---|---|---|---|---|---|
| | SST (°C) | $CO_2$ (ppm) | SST (°C) | $CO_2$ (ppm) | Coherency | (°) | kyr |
| 100 kyr | 1.0 | 28 | 1.0 | 28 | 0.98 | 9 ± 8 | 3 ± 2 |
| 41 kyr | 0.8 | 15 | 0.7 | 14 | 0.94 | 11 ± 14 | 1 ± 2 |

## 3. Deriving equilibrium tropical climate sensitivity estimates from the paleoclimatic data

### a. *Estimating climate sensitivity by linear regression*

Coupled ocean–atmosphere GCMs indicate a 500-yr or longer time interval required for climate to reach full equilibrium with a new forcing level (Cubasch et al. 2001). Because each 2-cm-long sediment interval in core TR163-19 integrates ~660 yr, the reconstructed SSTs approximate an equilibrium tropical climate value for that particular atmospheric $CO_2$ level. Therefore, it is possible to estimate equilibrium climate sensitivity by directly comparing tropical SSTs to known estimates of greenhouse forcing.

Direct point-to-point comparison of SSTs and $CO_2$ levels is achieved by interpolating both records to a constant 2-kyr sample interval, which is slightly greater than the average resolution of both records. Because the radiative effect of $CO_2$ is first order, changes in tropical temperature are compared to the calculated climate forcing using the expression (Ramaswamy et al. 2001)

$$\Delta F = 4.841 \ln(C/C_0) + 0.0906(\sqrt{C} - \sqrt{C_0}), \quad (1)$$

where $C$ is the atmospheric $CO_2$ value at a particular time and $C_0$ the 280 ppmv preindustrial $CO_2$ level. Because methane also varies on glacial–interglacial time scales (Petit et al. 1999), its effect, which is on average ~13% of the total forcing, has been included in the climate forcing calculation using the expression (Ramaswamy et al. 2001)

$$\begin{aligned}\Delta F = {} & 0.036 \ln(\sqrt{M} - \sqrt{M_0}) \\ & - 0.47 \ln[1 + 2.01 \times 10^{-5}(MN_0)^{0.75} \\ & \qquad + 5.31 \times 10^{-15} M(MN_0)^{1.52}] \\ & - 0.47 \ln[1 + 2.01 \times 10^{-5}(M_0N_0)^{0.75} \\ & \qquad + 5.31 \times 10^{-15} M_0(M_0N_0)^{1.52}], \quad (2)\end{aligned}$$

where $M$ is the atmospheric $CH_4$ value at a particular time, $M_0$ the 700 ppbv preindustrial $CH_4$ level, and $N_0$ the 275 ppbv preindustrial $N_2O$ level (past changes in $N_2O$ are ignored because of their relatively small radiative influence and the lack of a continuous record for the time period of interest).

The correlation between temperature change and total radiative forcing has a slope of 1.4° ± 0.1°C (W m$^{-2}$)$^{-1}$ and an intercept of 0.5° ± 0.2°C [errors are 95% confidence interval (CI); Fig. 5]. The positive temperature intercept is a consequence of SSTs warmer than modern conditions during discrete time spans of previous interglacials with similar $CO_2$ levels: MIS 5, 119–131 kyr BP; MIS 7, 233–245 kyr BP; and MIS 9, 325–335 kyr BP. This observation cannot readily be explained by greenhouse forcing alone, and either reflects another climate influence or inaccuracies in the data and/or time scale. One hypothesis to explain these warmer interglacial temperatures is that they reflect reduced ice volume during these time intervals; available data, however, do not suggest a significantly higher-than-modern sea level during peak MIS 5, 7, and 9 (Shackleton 2000; Lea et al. 2002; Waelbroeck et al. 2002; Siddall et al. 2003). An alternative hypothesis (N. Shackleton 2003, personal communication) is that the additional SST amplitude at 100 kyr that cannot be attributed to greenhouse forcing reflects the influence of the eccentricity envelope on precession. In this case, spectral analysis would extract both the 100-kyr SST amplitude due to greenhouse forcing as well as to the eccentricity envelope. The fact that the ratio of coherent amplitude of SST and climate forcing at 100 kyr only, which are 1.0°C and 0.7 W m$^{-2}$, respectively, is 1.4°C (W m$^{-2}$)$^{-1}$, identical to the value of $\lambda$ derived by fitting the discrete points (Fig. 5), is consistent with this hypothesis.

A convenient solution to the problem of a positive intercept is to force the regression through zero (i.e., assume $\Delta T$ is zero for a greenhouse forcing of zero), which yields a slope, equivalent to the climate sensitivity parameter $\lambda$ (Ramaswamy et al. 2001), of 1.10° ± 0.07°C (W m$^{-2}$)$^{-1}$ (errors are 95% CI of the slope; Fig. 5). Forcing the regression through zero yields the minimum value of $\lambda$, but with the consequence of re-

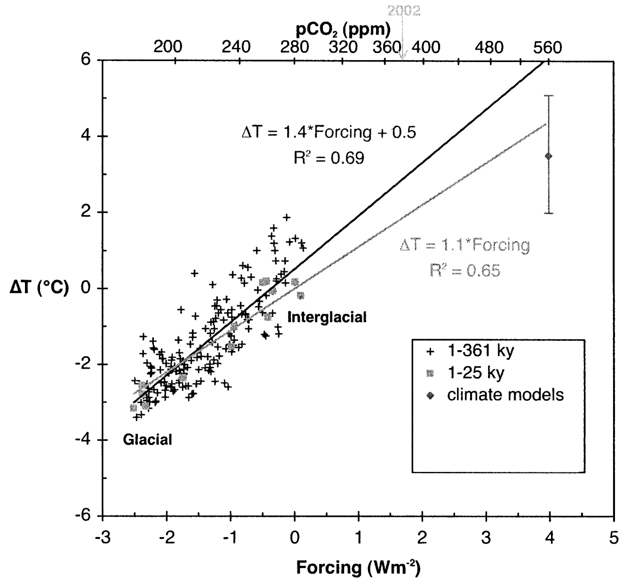


FIG. 5. Comparison of temperature differences ($\Delta T$), relative to modern sea surface temperature, at the tropical Pacific site, with greenhouse gas forcing calculated from Vostok $CO_2$ and $CH_4$ levels. The forcing is computed from standard radiative forcing relationships (Ramaswamy et al. 2001) relative to preindustrial levels of 280 ppm for $CO_2$ and 700 ppb for $CH_4$; $N_2O$ was assumed constant at 280 ppb (Raynaud et al. 2000). For reference, equivalent atmospheric $CO_2$ values are shown on the top axis. Both datasets have been interpolated to a constant sampling interval of 2 kyr, which is the average resolution of the tropical Pacific site (Lea et al. 2000). Regression through the data yields a slope, equivalent to the climate sensitivity parameter $\lambda$, of 1.4° ± 0.1°C (W m$^{-2}$)$^{-1}$ and the intercept of 0.5° ± 0.2°C ($r^2$ = 0.69; errors are 95% CI). Forcing the regression through zero yields a slope of 1.10° ± 0.07°C (W m$^{-2}$)$^{-1}$ ($r^2$ = 0.64). For the 4 W m$^{-2}$ forcing associated with a doubling of atmospheric $CO_2$, these values of $\lambda$ equate to a 4.4°–5.6°C warming.

TABLE 2. Results from the multivariate (MV) analysis, shown as % variance in the Cocos Ridge SST record explained by the various fits. ΣGHF calculated from the sum of $CO_2$ and $CH_4$ (see text). Insolation calculated using Paillard et al. (1996), with fall 2°N insolation integrated from Aug through Sep. The relative weighting for each factor in the multivariate fits was determined in a stepwise fashion, progressing from the strongest to the weakest weighting. Individual fits were determined by linear regression of the evaluated forcing against SST. All records (1–361 kyr BP) were interpolated to 2-kyr spacing.

| | Vostok ΣGHF | Jul 65°N insolation (lagged 2 kyr) | Fall 2°N insolation | Ice volume (Shackleton 2000) | Vostok ln(dust) | Tot |
|---|---|---|---|---|---|---|
| Best MV fit | 69% | 9% | 1% | 0% | 0% | 79% |
| Best MV fit without ΣGHF | — | 12% | 1% | 2% | 44% | 59% |
| Individual fits | 69% | 21% | 5% | 30% | 44% | — |

ducing the correlation between ΣGHF and SST, from $r^2$ of 0.69 to 0.63. The minimum slope can also be calculated by eliminating the warm points that create the positive intercept, in which case $\lambda = 1.1° \pm 0.1°C$ $(W\ m^{-2})^{-1}$, statistically identical to the prior approach.

### *b. Estimating climate sensitivity by multivariate regression*

A potential weakness of the approach just described is that it does not take into account other factors that could influence tropical SST, such as ice volume or atmospheric dust (Claquin et al. 2002). An alternative approach (Genthon et al. 1987) is to use multivariate regression to derive relative weightings for each of the hypothesized independent factors that control SST variations at the Cocos Ridge site. The following inputs were used for multivariate analysis: total greenhouse forcing (ΣGHF), local insolation (annual, spring, and fall), Northern Hemisphere (NH) insolation (July, 65°N), two different records of ice volume (Shackleton 2000; Lea et al. 2002), and records of Vostok ice core dust (both linear and ln; Petit et al. 1999). These records are not completely independent of each other (Bender 2003), but multivariate analysis provides a starting point to evaluate the relative weighting for each factor in the observed SST record.

In all of the multivariate analyses, ΣGHF accounts for the dominant signal, and its weighting always exceeds all of the other tested factors (Table 2). The best fit is achieved (79% variance explained) when July 65°N insolation, lagged by 2 kyr, is used as an ice volume proxy. For this fit, ΣGHF explains 69%, July 65°N insolation (lagged 2 kyr) explains 9%, and local fall insolation explains 1% of the variance. If ΣGHF is excluded from the analysis, up to 59% of the variance in SST can be explained by a combination of Vostok ln(dust), July insolation at 65°N, ice volume, and fall insolation at 2°N (Table 2). But because the Vostok ln(dust) and $CO_2$ records have a strong linear relationship (Bender 2003), it is difficult to assign any significance to this finding. Judged on an individual basis, ΣGHF explains 69% of the SST variance, Vostok ln(dust) explains 44%, the Shackleton (2000) ice volume record explains 30%, and July 65°N insolation (lagged 2 kyr) explains 21% (Table 2). The significant amount of variance explained by ΣGHF, dust, and ice volume in the individual fits indicates that it may be difficult to statistically separate the effects of these three forcings. For this reason, the dominance of ΣGHF in the multivariate effect does not rule out a potentially important role for the other forcings.

Examination of the relative weightings of each factor for the best multivariate fit demonstrates the dominance of ΣGHF in explaining the SST observations at the Cocos Ridge site (Fig. 6a). The effect of July 65°N insolation is significant, however, especially in explaining the observed SST minima in MIS 5.4 (5d) and MIS 7.4. Reconstructed SST, using the multivariate weighting, provides a good fit to the observed data (Fig. 6b), with the chief discrepancies occurring in MIS 5e (observed SST is too warm) and MIS 8.

The key finding from the multivariate analysis is that the value of climate sensitivity parameter $\lambda$ for the best fit is $1.3° \pm 0.1°C$ $(W\ m^{-2})^{-1}$, which falls within the range of the simple linear fit and the fit forced through zero (see above). Substitutions [such as using the Shackleton (2000) ice volume record instead of the July 65°N insolation] do not significantly affect this finding.

## 4. Implications for global climate sensitivity

For an assumed radiative effect of 4 W $m^{-2}$ for $CO_2$ doubling (Ramaswamy et al. 2001), the minimum value of $\lambda$ [$1.10° \pm 0.07°C$ $(W\ m^{-2})^{-1}$] implies a $4.4° \pm 0.9°C$ change in tropical SST [error calculated from the standard error (SE) of the regression and the estimated error of the SSTs (Lea et al. 2000)]. The maximum value of $\lambda$ [$1.4° \pm 0.1°C$ $(W\ m^{-2})^{-1}$] implies a $5.6° \pm 0.9°C$ increase for doubled $CO_2$. A regression through the data since the LGM only (1–25 kyr BP), which is the best constrained time interval from a chronological viewpoint, yields a sensitivity of $4.8° \pm 0.8°C$. The value of $\lambda$ derived from the best-fit multivariate analysis, $1.28° \pm 0.12°C$ $(W\ m^{-2})^{-1}$, yields a temperature response to doubled $CO_2$ of $5.1° \pm 0.8°C$. This value is probably the most representative estimate that can be derived from the data in this study.

It is instructive to compare this value to values derived by direct comparison of the Vostok D/H air temperature proxy and atmospheric $CO_2$ records (Petit et al. 1999). A multivariate study of the Vostok data (0–

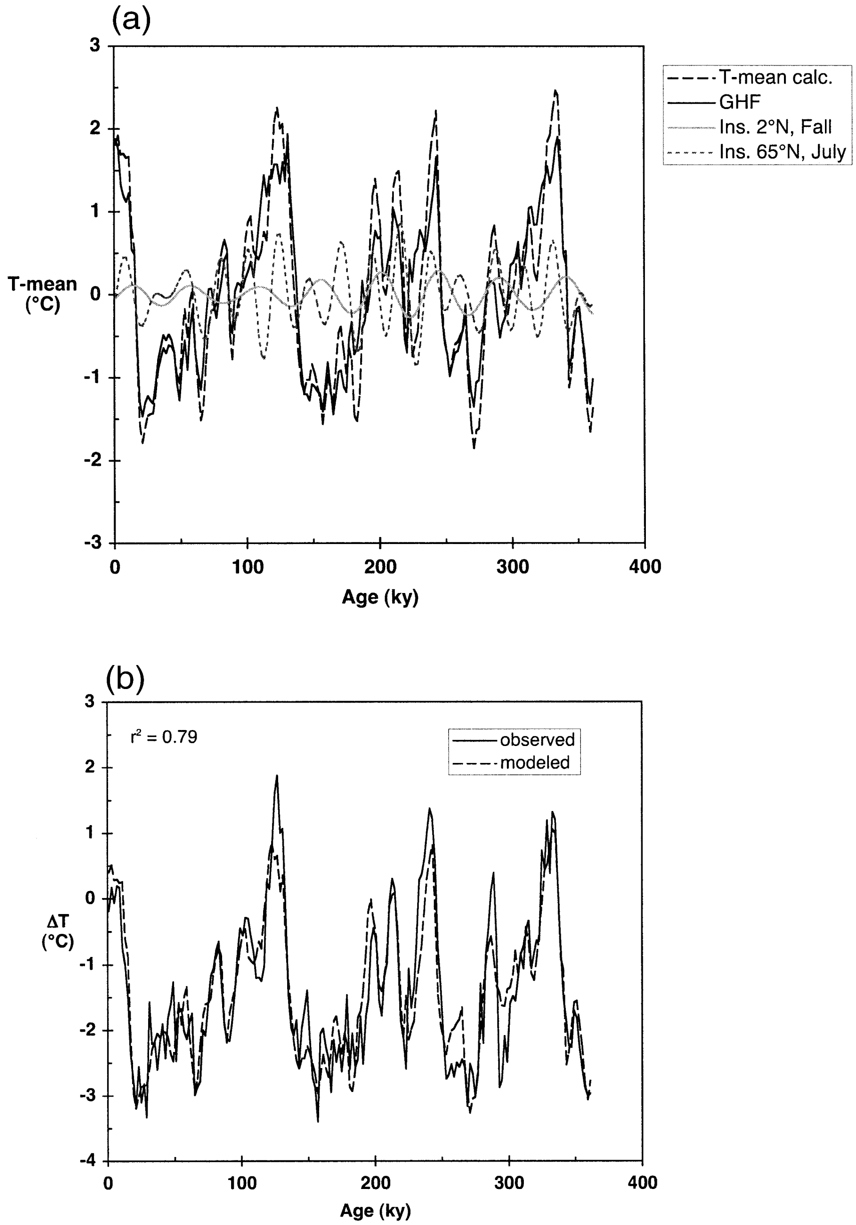


FIG. 6. Results from the multivariate analysis. (a) The relative weighting of the three climate forcings, $\Sigma$GHF, Jul 65°N insolation, lagged 2 kyr, and 2°N fall insolation (Paillard et al. 1996), that provide the best fit to the observed Cocos Ridge SST record (see Table 2 for the individual weightings). The temperature change that can be attributed to each factor is shown, along with the sum of the three influences (*T* mean calc). Temperature changes are shown as differences from the mean. (b) The modeled temperature change compared to the observed temperature change. The sum of three climate forcings explains 79% of the total variance in the SST record. Temperature changes ($\Delta T$) shown in differences from present SST.

160 kyr BP) found that for the majority of fits $CO_2$ variations accounted for 70%–85% of the air temperature change at Vostok, with amplification factors $f$ [i.e., ratio of observed temperature change to temperature change required to restore radiative equilibrium (Hansen et al. 1984)] of 11–14 (Genthon et al. 1987). This range of amplifications is equivalent to a local climate sensitivity of $\sim$3°–4°C (W m$^{-2}$)$^{-1}$. A subsequent study of the same dataset (Lorius et al. 1990) reported a much smaller amplification factor, $f \sim 3$, which is equivalent to a local climate sensitivity of $\sim$1°C (W m$^{-2}$)$^{-1}$. An alternative way to calculate climate sensitivity from the Vostok data is to ratio the 100-kyr amplitude of the D/H record that is coherent with $\Sigma$GHF, which is 13‰, equivalent to an air temperature change of $\sim$2°C (Jouzel et al. 2003), to the coherent 100-kyr amplitude in $\Sigma$GHF

as determined in this study, which is 0.7 W $m^{-2}$. This approach yields a local climate sensitivity of $\sim$3°C (W $m^{-2}$)$^{-1}$, in agreement with the Genthon et al. (1987) result. The polar amplification of greenhouse gas forcing relative to the Tropics using the approach is $\sim$2, as evaluated by taking the ratio of the coherent 100-kyr amplitude of the Antarctic air temperature record (2°C) to the coherent 100-kyr amplitude of the tropical SST record (1°C).

For comparison to the paleoclimatic approach, 15 climate models indicate a global mean equilibrium climate sensitivity of 3.5°C, with a range of 2.0°–5.1°C (Cubasch et al. 2001; Fig. 5). The tropical climate sensitivity from this study, 4.4°–5.6°C, is at the high end of the model-derived global climate sensitivity results. Because virtually all equilibrium climate simulations indicate that tropical temperature changes are smaller than the global mean, the global sensitivity implied by this study might lie above the IPCC range. The tropical climate sensitivity derived in this study is also larger than suggested by model simulations of the LGM climate (Hewitt and Mitchell 1997). This difference appears to be a consequence of the smaller effect that glacial $CO_2$ reduction has on tropical ocean temperatures in these models, which is typically 1°–2°C (Hewitt and Mitchell 1997; Ganopolski et al. 1998; Weaver et al. 1998; Broccoli 2000; Hewitt et al. 2003).

One factor that could explain the higher climate sensitivity values derived in this study relative to climate models is a strong cooling effect on the Tropics, via ocean and atmospheric heat transport, of mid- and high-latitude continental ice sheets. Modeling studies demonstrate that the influence of continental ice (i.e., ice volume) is vastly diminished in the Southern Hemisphere (Manabe and Broccoli 1985; Broccoli and Manabe 1987). Therefore, an independent way to evaluate ice volume influence is to compare glacial cooling from Southern Hemisphere warm pool sites. Two sites are presently available: 1) a site in the Makassar Straits (4°41.3′S, 117°54.17′E), which records a cooling of 3.3° $\pm$ 0.6°C during the LGM and 4.1° $\pm$ 0.6°C during the previous glacial maximum (MIS 6; Visser et al. 2003); and 2) a site in the Timor Trough (13°8.86′S, 121°34.69′E), which records a cooling of 2.8° $\pm$ 0.5°C during the LGM and 3.9° $\pm$ 0.8°C during MIS 6 (Opdyke et al. 2001). For comparison, the coolings observed at the Cocos Ridge site used in this study are 2.6° $\pm$ 0.8°C for the LGM and 3.9° $\pm$ 0.3°C for MIS 6. The similar level of cooling observed at all three sites supports the hypothesis that the dominant forcing is the glacial reduction in atmospheric greenhouse gases.

## 5. Limitations and future refinement of the approach

To what extent should the paleoclimatic results for equilibrium climate sensitivity be taken as representative of possible future changes? The advantage of the paleoclimatic approach is that it provides a series of equilibrium climate sensitivity experiments over a range of atmospheric $CO_2$ levels. It also involves no model assumptions other than the hypothesis that changes in tropical ocean temperatures are dominantly forced by varying $CO_2$ levels. A disadvantage of the paleoclimatic approach is that other climatic influences, such as the size and distribution of the glacial ice sheets, surface albedo, insolation changes, and concomitant changes in ocean and atmospheric circulation all could have played a role in determining the relationship between tropical climate and $CO_2$ in the past in such a way that the past pattern does not apply to the future. But, growing paleoclimatic evidence that the timing of tropical temperature changes leads ice volume changes by several thousand years or more (Pisias and Mix 1997; Schneider et al. 1999; Lea et al. 2000; Stott et al. 2002; Visser et al. 2003), suggests that the effect of continental ice on tropical SSTs is likely to be a very slow feedback, at least compared to the rapid equilibration time of the tropical mixed layer. The other influences are harder to separate, but GCM results do not uniformly identify any of them as likely to be an important component of the glacial climate forcing in the Tropics (Hewitt and Mitchell 1997; Ganopolski et al. 1998; Weaver et al. 1998; Broccoli 2000). The more recent suggestion that atmospheric dust loading could exert a considerable influence (up to 3 W $m^{-2}$ locally) on tropical climate adds another factor with potentially the same magnitude as greenhouse forcing (Claquin et al. 2002). Dust influence, however, has a very strong spatial pattern mediated by source regions (Claquin et al. 2002), and the presently available paleo-SST data indicate a relatively uniform glacial cooling of $\sim$3°C for the tropical warm pools, which is probably not consistent with the pattern expected from a strong dust influence.

There are a number of possibilities for future work that can be pursued to evaluate the results of this study. Ideally, any calibration of tropical SST versus atmospheric $CO_2$ would not rely on a single record, because any particular site is likely to reflect some combination of the influence of changing greenhouse forcing as well as local climate factors. The availability of a stacked record of tropical SST combining records from the warm pools of all three ocean basins should average out local effects and improve tropical climate sensitivity determination from paleoclimate data. An additional uncertainty stems from the absolute magnitude of SST change reconstructed from the foraminiferal Mg/Ca proxy; the largest uncertainty in paleotemperature analysis appears to be related to potential changes in foraminiferal preservation in the past due to postdepositional changes (Lea et al. 2000; Dekens et al. 2002; Rosenthal and Lohmann 2002), but secondary environmental influences, such as changing pH, might introduce additional uncertainty (Lea et al. 1999). By comparing and combining records from different basins with different preservational histories, it should be possible to reduce the uncertainty associated with postdepositional changes. From the viewpoint of climate process, the biggest uncertainty lies in the assumption that atmospheric $CO_2$ alone drives tropical SST change at the 100-kyr period. Modeling runs, which could separate the response of tropical temperatures to potential glacial climate forcing other than $CO_2$, would be useful in this regard. In addition, it would be useful if model runs could be used to test the assumption that climate sensitivity determined from a colder past can be applied symmetrically to a warmer future.

## 6. Conclusions

Tropical SST variations spanning the last half million years are dominated by the 100 000-yr cycle. A proposed hypothesis to explain this observation is that varying atmospheric carbon dioxide abundances, which are a likely source of 100-kyr variability in paleoclimate records (Shackleton 2000), are directly forcing tropical climate on long orbital time scales. This hypothesis is tested by comparing a 360-kyr tropical SST record from the Cocos Ridge, north of the Galapagos Islands, with the Vostok atmospheric $CO_2$ record. The comparison indicates that the two records have a strong overall correspondence ($r = 0.84$) and are highly coherent with no phase offset at 100 and 41 kyr, supporting a greenhouse gas mechanism as the dominant control of tropical climate change. The phasing provides an explanation for the observed lead of tropical SST relative to ice volume (Lea et al. 2000), because atmospheric $CO_2$ changes early in the climate cycle (Petit et al. 1999; Shackleton 2000). Integrating the available observations suggests that atmospheric $CO_2$, Antarctic temperature, and tropical temperature all changed early in the climate cycle relative to ice volume.

Tropical climate sensitivity is computed from the Cocos Ridge comparison by directly comparing the amplitude of SST variations to that of radiative forcing computed from varying atmospheric $CO_2$ and $CH_4$. The computation indicates a climate sensitivity $\lambda$ of 1.1°–1.4°C (W m$^{-2}$)$^{-1}$ [error estimated at ± 0.1°C (W m$^{-2}$)$^{-1}$]. Using this range, a 4 W m$^{-2}$ radiative forcing associated with a doubling of atmospheric $CO_2$ would equate to a 4.4°–5.6°C increase in temperature (total errors estimated at ±1°C), generally above the currently accepted Intergovernmental Panel on Climate Change (IPCC) range of 1.5°–4.5°C from GCMs (Cubasch et al. 2001). The degree to which climate sensitivities determined from paleoclimatic data apply to future warming depends on how differently climate feedbacks operated in the past compared to the future.

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