4.5 MODELLING AND INTERPRETATION OF OXYGEN ISOTOPE RECORDS OF TROPICAL CLIMATE VARIABILITY

Josephine Brown* and Ian Simmonds
University of Melbourne
Victoria, Australia

1. INTRODUCTION

Information about past tropical climate variability is recorded in oxygen isotope ratios in coral and foraminiferal carbonate as well as tropical ice cores. The interpretation of isotopic climate records requires assumptions about the application of present day isotope-climate relationships to past climate. These are tested here using an atmospheric climate model fitted with a stable water isotope tracer scheme.

A 50-year simulation of present day climate is compared with available isotopic observations in order to validate the model. The climate of the Last Glacial Maximum is simulated, and the modelled tropical isotopic distribution in precipitation and the surface ocean discussed in relation to paleoclimate reconstruction.

2. MODEL VALIDATION FOR PRESENT DAY CLIMATE

The Melbourne University General Circulation Model (MUGCM) is fitted with an isotopic water tracer scheme allowing the ratio of oxygen isotopes in vapour and precipitation to be calculated. The atmospheric isotopic tracer scheme has been found to accurately reproduce the observed seasonal and annual distribution of isotopes in precipitation (Noone and Simmonds, 2002).

The model also includes a one-dimensional surface ocean isotopic scheme that calculates the isotopic ratio from the precipitation and evaporation at each ocean grid point:

$$\partial W/\partial t = P - E + k\,(h - W) \qquad (1)$$

$$\partial W_i/\partial t = P_i - E_i + k\,(h\,R_{SMOW} - W_i) \qquad (2)$$

where W and W_i are the surface (mixed layer) depths for $H_2{}^{16}O$ and $H_2{}^{18}O$ respectively, and P, P_i, E, E_i are the precipitation and evaporation terms, h is the mean mixed layer depth, k is the damping timescale and R_{SMOW} is the mean ocean isotopic ratio. This scheme enables the surface ocean ratio $R = W/W_i$ to be calculated interactively, rather than assuming a constant surface ocean ratio as previous isotopic modelling studies have done (e.g. Jouzel *et al.*, 1987; Hoffmann *et al.*, 1998).

A simulation of present day (1950-1999) climate was carried out with MUGCM forced with GISST2.3b sea surface temperature (SST) and sea ice data (Rayner *et al.*, 1996). The model reproduces interannual climate variability in the tropics, such as El Niño-Southern Oscillation (ENSO) and decadal variability associated with the Pacific Decadal Oscillation (PDO). The isotopic ratio varies with the amount of precipitation in this region following the well know isotopic "amount effect" (Dansgaard, 1964) whereby greater precipitation amount is accompanied by rain-out of the heavier isotopic species, leaving the remaining vapour and hence precipitation increasingly isotopically light.

The correlations between the modelled Southern Oscillation Index (SOI) of ENSO activity and monthly anomalies of precipitation amount and $\delta^{18}O$* of precipitation (Figures 1a and 1b) show that isotopic ratio and precipitation are inversely related to ENSO activity. In the central Pacific increased precipitation during El Niño events is associated with more negative $\delta^{18}O$ values. The reverse occurs over the west Pacific. A similar but weaker relationship is seen between the PDO index and $\delta^{18}O$ (not shown).

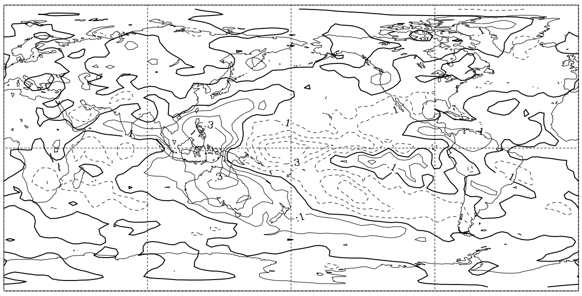

Figure 1a: Correlation between precipitation and SOI (contour interval = 0.1)

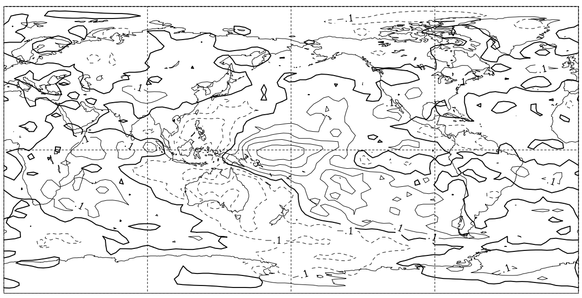

Figure 1b: Correlation between $\delta^{18}O$ and SOI (contour interval = 0.1)

The isotopic signal in the surface ocean was also analysed on seasonal and interannual timescales. The average surface ocean isotopic distribution was

* *Corresponding author address*: Josephine Brown, School of Earth Sciences, University of Melbourne, VIC 3010, Australia; email: jbrown@earthsci.unimelb.edu.au

* $\delta^{18}O = [(^{18}O/^{16}O)/R_{SMOW} - 1] \times 1000‰$ where R_{SMOW} is the ratio $^{18}O/^{16}O$ in Standard Mean Ocean Water.

compared with observations from Schmidt *et al.* (1999). The model was found to capture the large-scale tropical isotopic distribution, with enrichment in areas of high evaporation and depletion in regions of high precipitation (Figure 2).

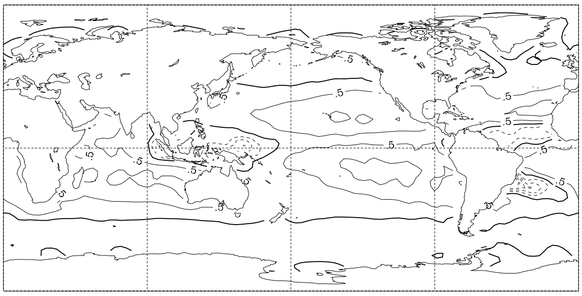

Figure 2: Surface ocean $\delta^{18}O$ for present day (contour interval = 0.5‰)

The interannual variability in the modelled surface ocean isotopic ratio was also compared with observations. The strongest signal was associated with ENSO, showing a positive correlation between SOI and $\delta^{18}O$ in the central Pacific, as well as a signal in the eastern equatorial Pacific associated with SST-related changes in evaporation.

An additional validation of the model is to compare coral isotopic records from the last fifty years with ratios calculated offline from modelled SST and surface ocean $\delta^{18}O$ values. The oxygen isotope ratio in coral carbonate depends on both the local ocean isotopic ratio and a biological SST-dependent fractionation:

$$\delta^{18}O_{coral} = \delta^{18}O_{ocean} + a\ (SST) + b \qquad (3)$$

The isotopic ratio of modelled "coral" was calculated for the modelled surface ocean $\delta^{18}O$ and SST using empirical SST/isotope coefficients for *Porites* coral (a = –0.2‰/°C, b = 0.45‰) (Juillet-Leclerc and Schmidt, 2001) and the result compared with coral samples from selected locations. An additional correction of –0.27‰ due to conversion from coral PDB standard to ocean RSMOW standard was required.

In regions where SST dominates the coral isotopic signal the modelled and observed coral $\delta^{18}O$ are correlated, with warmer SSTs producing more negative $\delta^{18}O$ values. At Amdee, New Caledonia (22S, 176E) in the southwest Pacific, coral $\delta^{18}O$ has been sampled at quarterly intervals from 1950-1990 (Quinn *et al.*, 1998). While the correlation between observed and modelled coral $\delta^{18}O$ is weak (r = +0.2), both records vary with local SST (Figure 3a).

In regions such as the central Pacific where precipitation variability alters the surface ocean isotopic ratio during ENSO events, the model was also able to reproduce the observed anomalies of coral $\delta^{18}O$. The monthly coral isotopic record at Kiritimati (2N, 157W) from 1950-1993 (Evans *et al.,* 1993) was compared with the modelled coral $\delta^{18}O$ (r = 0.74). Both isotopic ratios are strongly correlated with SOI (r = 0.82 for modelled coral $\delta^{18}O$, r = 0.77 for observed coral $\delta^{18}O$) as seen in Figure 3b, indicating that the model reproduces the local precipitation variability and hence surface ocean $\delta^{18}O$ variability associated with observed ENSO events.

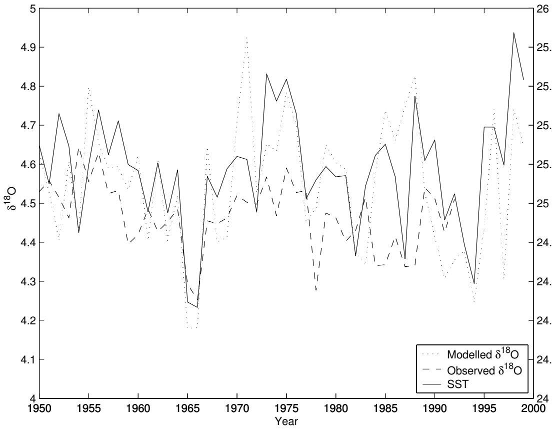

Figure 3a: New Caledonia: modelled and observed coral $\delta^{18}O$ and SST ($\delta^{18}O$ reversed for comparison)

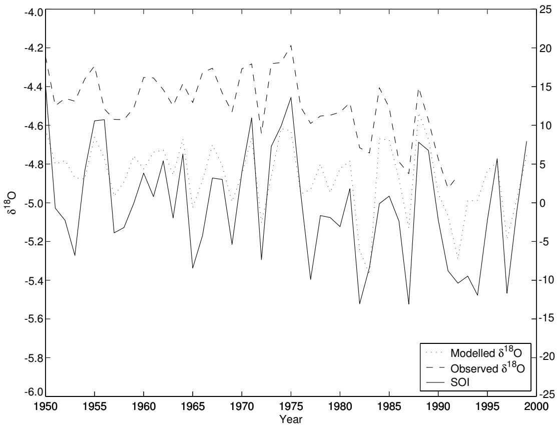

Figure 3b: Kiritimati: modelled and observed coral $\delta^{18}O$ and SOI

In summary, the model reproduces the average tropical isotopic precipitation and surface ocean distribution, and important features of the interannual climate variability such as ENSO. The ability of the model to capture the main features of the seasonal and interannual $\delta^{18}O$ distribution in the tropics confirms that it can be confidently used to investigate sensitivity to glacial climate change.

3. SIMULATION OF THE LAST GLACIAL MAXIMUM

A simulation of the Last Glacial Maximum (21 kyr BP) climate was carried out using CLIMAP SSTs and altered sea ice coverage, sea level and atmospheric CO_2 concentrations. The surface ocean isotopic ratio was enriched by +1.6‰ to reflect the change in sea level leading to an increase in the global ocean $\delta^{18}O$ value.

The resulting precipitation $\delta^{18}O$ and surface ocean $\delta^{18}O$ anomalies (LGM – present day) are shown in Figures 4a and 4b. In mid- and high-latitudes, the $\delta^{18}O$ of precipitation decreases with cooling at the LGM, while the tropical signal follows changes in the amount of precipitation. Over most of the tropics the anomaly is positive, due to the surface ocean enrichment as well as a decrease in the amount of precipitation. In the southeast Pacific, the modelled precipitation amount increases during the LGM and hence the precipitation $\delta^{18}O$ anomaly is negative.

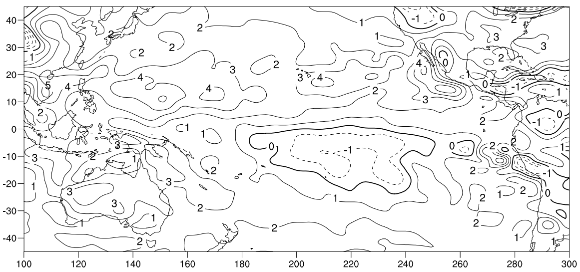

Figure 4a: Precipitation $\delta^{18}O$ (LGM – present) (contour interval = 1‰)

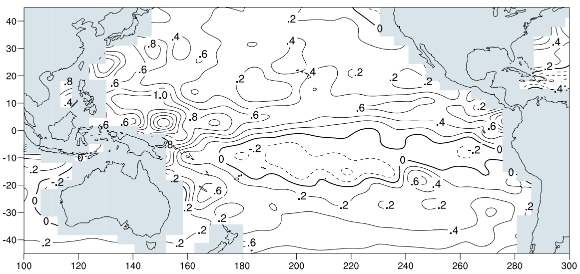

Figure 4b: Surface ocean $\delta^{18}O$ (LGM – present) (contour interval = 0.2‰)

The surface ocean signal reflects changes in the balance between precipitation and evaporation. The modelled surface ocean $\delta^{18}O$ anomalies are shown with the +1.6‰ global ocean enrichment removed. The largest anomalies are seen in the west Pacific where decreased precipitation produces positive ocean $\delta^{18}O$ anomalies and in the central Pacific where increased precipitation and decreased evaporation lead to negative ocean $\delta^{18}O$ anomalies.

The change in surface ocean $\delta^{18}O$ values under LGM climate conditions indicates that isotopic records such as fossil coral and planktonic foraminifera cannot be simply interpreted as records of changes in SST. In some regions, local ocean $\delta^{18}O$ variability will dominate where the precipitation minus evaporation balance is altered, while changes in seasonality of precipitation may also be important.

The relative magnitude of these two influences on coral (and foraminiferal) isotopic records is illustrated by comparing the component of coral $\delta^{18}O$ change due to local ocean water $\delta^{18}O$ change (Figure 4b) with that from SST change (assuming –0.2‰/℃), as shown in Figure 5.

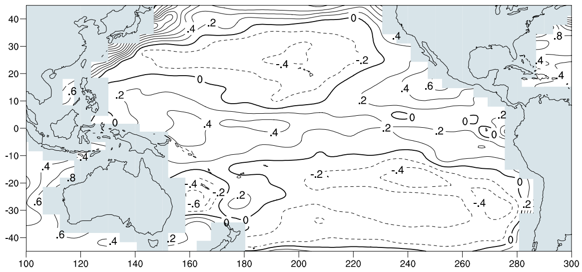

Figure 5: –0.2‰/ºC x SST (LGM – present) (contour interval = 0.2‰)

The magnitude of the modelled SST-induced isotopic signal is small in the west Pacific, while the reduced precipitation in this region leads to a larger change in the local ocean $\delta^{18}O$. In the central and east Pacific, the magnitudes of SST and ocean $\delta^{18}O$ components of the isotopic record are similar, indicating that comparison with other independent records (such as Sr/Ca ratio) is required to constrain the reconstructed SST change in these regions.

REFERENCES

Dansgaard, W., 1964: Stable isotopes in precipitation, *Tellus*, **16**, 436-468.

Evans, M.N., R.G. Fairbanks and J.L. Rubenstone, 1998: A proxy index of ENSO teleconnections, *Nature*, **394**, 732-733.

Jouzel, J., G.L. Russell, R.J.Suozzo, R.F. Koster, J.W.C. White and W.S. Broecker, 1987: Simulation of HDO and $H_2{}^{18}O$ atmospheric cycles using the NASA GISS general circulation model: The seasonal cycle for present-day conditions. *J. Geophys. Res.*, **92**, 14739–14760.

Juillet-Leclerc, A. and G.A. Schmidt, 2001: A calibration of the oxygen isotope paleothermometer of coral aragonite from Porites. *Geophys. Res. Lett.,* **28**, 4135-4138.

Hoffmann, G., M. Werner and M. Heimann, 1998: Water isotope module of the ECHAM atmospheric general circulation model: A study on timescales from days to several years. *J. Geophys Res.*, **103**, 16871-16896.

Noone, D, and I. Simmonds, 2002: Associations between $\delta^{18}O$ of water and climate parameters in a simulation of atmospheric circulation for 1979-1995. *J. Climate*, **15**, 3150-3169.

Quinn, T.S. *et al.,* 1998: A multicentury coral stable isotope record from a New Caledonia coral. *Paleoceanography*, **13,** 412-426

Rayner, N.A. *et al.,* 1996. *Clim. Res. Tech. Note **74***, UK Meteorological Office, Bracknell.

Schmidt, G.A., G.R. Bigg and E.J. Rohling, 1999: “Global Seawater Oxygen-18 Database”. *http://www.giss.nasa.gov/data/O18data*