Location and hydrography of study area

Three deep-sea gravity core samples from Shatsky Rise, northwestern Pacific, were used for the present study (Figure 1). The cores were collected during the *R/V Hakurei-maru* cruise in 1995 as part of the North Pacific Carbon Cycle Study of the New Energy and Industrial Technology Development Organization. Initially, two cores were studied: NGC 102 (32°19.84' N and 157° 51.04' E) from a depth of 2,612 m; and NGC 108 (36° 63.85' N and 158° 20.90' E) from a depth of 3390 m. An additional core sample, NGC 106 (34 09.93'and 158 45.18' E; 3,713 m depth), located between NGC 102 and NGC 103, was included in the study primarily to constrain glacial southward extent of sub-polar frontal boundary.

Shatsky Rise is an irregularly shaped plateau rising above the mostly abyssal sea floor of the northwestern Pacific. It lies along the path of the eastward extension of the Kuroshio Current. The Kuroshio Current is the western boundary current in the North Pacific and its extension marks the northern limit of the Western North Pacific Water Mass (Bradshaw, 1959). Steep temperature, salinity, and nutrient gradients (Figures 2 and 3) characterize the broad transition region between the cold and nutrient rich sub-arctic waters and the warm and sterile central waters. Whereas the sub-artic waters is characterized by a permanent halocline, the central waters is characterized by a perament thermocline with a thick mixed layer (Thompson, 1981). The transition region is a site of intense mixing and is believed to be locus of North Pacific Intermediate Water formation (Reid, 1965). Thompson (1981) further subdivided the Western Pacific Central Mass into a transitional and a sub-tropical water masses based on planktic foraminiferal assemblages. The present position of the sub-artic frontal boundary is estimated to be between 40^0 and 45^0N latitude (Bradshaw, 1957; Thompson, 1981).

Beneath the mixed layer of the Kuroshio Extension is the North Pacific Intermediate Water which is centered at depths of 300 - 600 m and which can be traced to a depth of 1,000 m in low to mid latitudes (Reid, 1965). From a depth of 2,000 m to the bottom (Figures 4 a, b), the waters generally attain uniform properties characteristic of the Pacific Deep Waters. There is presently no deep water formation in the Pacific (Warren, 1983; Keigwin, 1987; Broecker, 1995) and the deep water found in the region is thought to originate from the south and is the result of a mixture of the Antarctic Bottom Water and the North Atlantic Deep Water (Millero, 1996). The CCD is estimated to be between 4,000 and 4,500 (Berger and Winterer, 1974) whereas the depth of the sedimentary lysocline has not been defined precisely. The depth of the foraminiferal lyscoline should, however, be shallower than 2,900 m (Vincent, 1975).

Methodology and Theoretical Background

Laboratory processing

Core samples NGC 102 and NGC 108 were sampled at 8 cm intervals. Core sample NGC 106 was sampled at 8 cm intervals for the first top 2 meters and at 16 cm intervals hence. The samples where disaggregated by soaking in warm (~50°C) water while gently stirring. After just a few minutes, they were washed clean through a 63-μm sieve in a gentle shower avoiding undue mechanical destruction of foraminiferal tests. Washed residues were then oven dried at ~60°C and weighed. Dried samples were then dry sieved through a 150-μm sieve and the fractions stored in separate containers. Individual > 150-μm fractions were split repeatedly in order to obtain a sample split that would yield approximately 300 whole planktic foraminiferal specimens. Using the same final split, the number of foraminiferal fragments were counted.

From a separate sample split, 100 specimens of *Globorotalia inflata* were picked for oxygen isotope analysis. Picked specimens were ultrasonically cleaned while immersed in methanol inside small glass vials and then oven dried. Approximately 1 mg of the prepared *G. inflata* specimens per sample were used in the analysis.

Dissolution indices

Two planktic foraminiferal indices of dissolution were used in the present study: Berger's (1981) foraminiferal dissolution index (FDX) and the relative proportion of fragments (FRAG). FDX is based on the relative proportion of dissolution resistant planktic foraminiferal species. This value can be calculated using the formula:

$$FDX = \Sigma_i^n (pi\ Ri) / \Sigma_i^n pi$$

where *pi* is the percentage of species *i* and R*i* is the rank of species *i*. Berger's (1981) ranking of foraminiferal species in the order of susceptibility to dissolution (Table 1) is here adapted. The relative proportion of fragments (FRAG) is given by the following formula:

$$FRAG = \frac{\#\text{ of foraminifer fragments}}{\#\text{ of foraminifer fragments} + \#\text{ of whole foraminifer test}}$$

Any part of a foraminiferal test with a size less than half of the inferred original size is here considered as a fragment (Williams *et al.*, 1985; Malmgren, 1987; LaMontagne *et al.*, 1996). In addition, these two indices are compared with planktic foraminiferal number (# of planktic foraminifera per gram sediment), and $CaCO_3$ content when data is available. The merits and limitations of the use of the different estimates of dissolution intensity have been sufficiently discussed by LaMontagne *et al.*(1996). Table 2 is a list of the commonly used indices to track dissolution intensity. Although benthic foraminiferal abundance is also used as an index of dissolution, the results of Ohkushi's (1998 MS) study of the same core (NGC 102) suggest that benthic foraminiferal number primarily responded to changes in primary paleoproductivity levels. As expressed by LaMontagne *et al.*(1996), confidence in the effectivity of the use of these individual proxies to exclusively track dissolution intensity increases when these parameters covary down-core. It should be emphasized that sensu strictu, it is the relative degree of foraminiferal dissolution which is here approximated. It can but is not necessarily proportional to carbonate content as the latter is subject not only to the effects of dissolution, but is also dependent on dilution and the rate of carbonate supply (Berger,1992; Karlin *et al.*, 1992; Haug *et al.*; 1995). Furthermore, using these two indices alone, we are not able to distinguish between changes in dissolution intensity as a result of changes in the rate of production of metabolic CO_2 within the sediment and changes in bottom water carbonate saturation.

Sea Surface Temperature Estimation

There are currently two planktic foraminiferal based methods used in the reconstruction of sea surface temperature: the I & K method introduced by Imbrie and Kipp (1971), and the Modern Analogue Technique (MAT) by Hutson

(1980). The two basic assumptions for both methods are: a) similar foraminiferal assemblages are produced by similar suites of environment conditions, and b) sea surface temperature is the primary controlling factor or correlated with environmental variables which determine foraminiferal variation. (Imbrie and Kipp, 1971; Prell, 1985; Pflaumann *et al.*, 1996; Ortiz and Mix, 1997). An intrinsic corollary of the first assumption is that species physiological adaptation does not necessarily change with time. It is for this reason that most sea surface temperature reconstructions based on planktic foraminifers are limited to the Pleistocene to Recent sequences; an interval representing the period when the modern species have since existed. There are, however, recent attempts to apply these methods to sequences as old as the Pliocene (*e.g.* Dowsett, H.J. 1991; Thompson *et al.*, 1994; Andersen, 1997). Several studies comparing the relative performance of the I &K method and MAT exist in recent literature (*e.g.* Prell, 1985; Andersen, 1997, Ortiz and Mix, 1997).

FP12-E Transfer Function

The foraminiferal transfer function FP12-E was developed by Thompson (1981) using 186 deep-sea piston core tops from the western North Pacific. Q-mode factor analysis delineated 6 faunal assemblages related to water masses: Polar/subpolar Assemblage, Gyre Margin Assemblage, Transitional Assemblage, Sub-tropical Assemblage, Tropical Dissolution Susceptible, and Tropical Dissolution Resistant Assemblage. These six assemblages represent more than 93% of the total variance of the data set. Using the procedure outlined by Imbrie and Kipp (1971), a linear transfer function relating the six varimax factors obtained from the Q-mode factor analysis to summer and winter temperatures was obtained. Standard errors of the two temperature estimates are 1.5^{0}C and 2.5^{0}C, respectively.

Table 3 is a list of the planktic foraminiferal species and maximum assemblage percentages used in development of the transfer function FP12-E. Ideally, when using paleoecological transfer functions, the maximum percent occurrence of each of the species in the sample in question should not be more than that of the calibration set. In cases where the species percentages are beyond the maximum species occurrence in the calibration set, no analog conditions exist and will result to low communalities. In such cases, the resulting temperature-estimates (Hutson, 1977; Sachs *et al.*, 1977, Andersen, 1997) are extrapolated values that are either over or under estimates. A comprehensive discussion of the recognition and treatment of no analog conditions in the use of transfer functions is found in Hutson (1986).

The varimax factor score matrix and the parameters of the transfer function FP-12E are given in tables 4 and 5, respectively. The normalized relative abundance of the individual planktic foraminiferal species are related to temperature using the following equation:

$$\begin{aligned} T\ est. = K \quad & + a(\Sigma_i^n UiF5i) + b(\Sigma_i^n UiF2i) + c(\Sigma_i^n UiF1i) \\ & + c(\Sigma_i^n\ UiF1i) + d(\Sigma_i^n UiF3i) + e(\Sigma_i^n\ UiF6i) \\ & + f(\Sigma_i^n\ UiF4i) \end{aligned}$$

Where:

T est. = estimated Sea Surface Temperature
K, a, b, c, d, e and f are regression coefficients;
Ui = normalized species abundance of species "i" in factor 1;
F1i = factor score of species "i" in factor 1;
F2i = factor score of species "i" in factor 2;
F3i, F4i, F5i, F6i as follows.

The advantages of using FP-12E over all other planktic foraminifral transfer functions in reconstructing sea surface temperature of the present study area are: a) the calibration data set used is limited to the western North Pacific and therefore more sensitive to the local ecology, and b) samples included in the calibration set cover a wide range of depths and not limited to well preserved sediments.

Modern Analog Techinque

Sea Surface Temperature estimates derived using the Modern Analoque Technique are based from direct comparison with modern day assemblages or analogs (hence the name).

From a database of core top assemblage with known evironmental parameters, assemblages/localities with the closest similarity to the sample with the unkown environmental parameter are chosen. There are several methods employed to express similarity or dissimilarity between assemblages in the modern analog technique (Overpeck *et al.*, 1985), but the square chord distance is the most common. SST is computed from the weighted average of the SST's of the data base samples with the highest similarity.

In the present study, MAT derived SST estimates were obtained using the software ANALOG (Schweitzer, 1997). The CLIMAP (1985) core top database (included in the program) was used and the weighted average of both summer and winter SSTs of the top 5 samples with the highest similarity (square chord distance) were adapted.

Lithology and chronology of cores

Lithologic description including down-core magnetic susceptiblity profiles, detailed color descriptions, and preliminary nannofossil age determinations can be found in the 1995 *R/V Hakurei-maru* cruise report (Nishimura and Kawahata, 1997). The three cores are composed predominantly of calcareous oozes punctuated by distinctly more siliceous intervals (Figure 5). Age model constructed by Kawahata *et al.* (*in press*) for NGC 102 is adapted in the study. For both NGC 102 and 108, the planktic foraminiferal species *Globorotalia inflata* was used for oxygen isotope analysis. Age model for NGC 108 were constructed by graphic correlation with SPECMAP (Figure 6) using the software Analyseries 1.0 (Paillard *et al.*, 1996). In addition, several biostratigraphic datum were useful as age control points. The last down-core occurrence of the warm water species *Globigerinoides ruber* pink in NGC 102 is consistent with the timing of its reported disappearance in the Pacific at approximately 122 kyr (Thompson *et al.*, 1979). This datum also coincides with abrupt change from predominantly right to left coiling of *Globorotalia truncatulinoides* reported in the northwestern Pacific (Xu *et al.*, 1995). The general absence of *G. ruber* pink and the rare occurrence of *G. truncatulinoides* in core NGC 106 and 108 preclude the use of these two biostratigraphic events at least for this more northern core. Down-core abundance patterns of *Neogloboquadrina pachyderma* left-coiling in core sequences at or near subpolar frontal boundary have been used extensively not only for correlation but also to track movement of the polar frontal boundary in the North Pacific (Thompson, 1977, 1981; Xu *et al.*, 1995). This is particularly useful for core NGC 108 and NGC 106 where *N. pachyderma* left-coiling down-core abundance patterns show distinct glacial-interglacial trend. Oxygen isotope analysis was also conducted for NGC 106 but the results have not be obtained as of writing. The age model for this core is constructed by multiple correlation of fragmentation and paleotemperature curve with the NGC 108 age/fragmentation model and SPECMAP, respectively (Figure 6b).