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Seasonal sea level change from TOPEX/Poseidon observation and thermal contribution

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Abstract. Seasonal steric sea-level change due to temperature variation in the mixing layer is assessed using space-measured sea-surface temperature data and historical in situ temperature measurements. The results are compared with TOPEX/Poseidon satellite altimeter measurement at different large spatial scales. It is indicated that thermal effect accounts for much of the observed seasonal variability, especially when averaging over zonal regions. Some regional seasonal patterns of sea-level anomalies in the tropical oceans are well represented by the thermal model prediction. Systematic differences are shown between TOPEX/Poseidon observation and thermal contribution at a 1–2 cm level. The potential causes for these differences are discussed, including water mass exchanges among the atmosphere, land, and oceans, and error sources in the steric result and geophysical corrections applied in TOPEX/Poseidon data.

Key words: TOPEX/Poseidon – Sea-level change – Thermal effect

1 Introduction

The TOPEX/Poseidon satellite altimeter has been monitoring global sea-surface height change every 10 days for over 6 years with an accuracy approaching 3–4 cm over a spatial scale of 20 km along the satellite ground track (Fu and Davidson 1995). The high spatial resolution and over 5 years' temporal coverage provide a broad band of information about sea-level variation and geostrophic ocean circulation. Observed sea-surface height is subject to change as a consequence of many effects, including ocean tides and solid earth tides, the pole tides, atmospheric pressure loading, steric effects introduced by temperature and salinity variations within the oceans, and water mass redistribution associated with the global hydrologic and cryospheric cycles.

After all standard geophysical corrections are applied to TOPEX/Poseidon data (Callahan 1993), there remain strong seasonal sea-level variations at various spatial scales with a mean annual variability of ±4–5 cm in the northern hemisphere and ±2–3 cm in the southern hemisphere. The maximum and minimum sea-surface heights in the northern hemisphere are in September and March, respectively, and the phase is reversed in the southern hemisphere (see Fig. 1). It is generally believed that these seasonal signals reflect the seasonal fluctuations of temperature over the oceans. Many investigations (Repert et al. 1985; White and Tai 1995; Chambers et al. 1997) show that sea-level variations are highly correlated with heat storage change. The inferred heat storage change using TOPEX/Poseidon-determined sea-level variations is in very good agreement with the heat storage change calculated from temperature profiles in certain basins (White and Tai 1995; Chambers et al. 1997). However, in terms of geodynamic applications of satellite altimeter data, people are more interested in large-scale steric and non-steric sea-level change on a global basis.

Steric sea-surface height stands for the portion of sea-level change due to density variation, which is introduced by temperature and salinity variation and is dominated by thermal effect. We neglect salinity effects in our steric estimation because of two considerations: the lack of reliable salinity data on worldwide ocean scales and the difficulties in separating the salinity variations due to ocean currents and salt convection from the salinity variations in the top of the mixing layer due to fresh water in-flux and out-flux associated with precipitation, evaporation and river discharge (in specific regions). The latter is caused by oceanic mass variation, which is actually part of the non-steric, or mass-related information. Steric sea-level change is a non-linear in-

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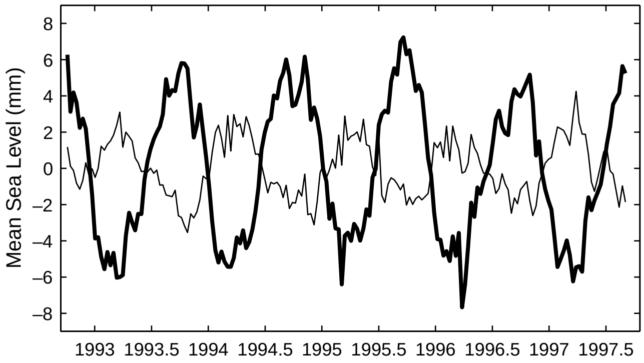

Fig. 1. The mean sea-surface anomalies in regions between 20 and 50 degrees in the northern (*thick curve*) and southern (*thin curve*) hemispheres, measured by TOPEX/Poseidon altimetry (cycles 2–183)

tegral of temperature variation, because thermal expansion coefficients depend on temperature and pressure, especially temperature (see Fig. 2). A better understanding of steric sea-level change leads to the potentials in separating mass variation within the ocean from observed sea-level anomalies, which is of great interest to many studies, including water mass transport within the global hydrological cycle, oceanic impact on the Earth rotational and gravitational change, and the validation of geophysical model performance in altimeter data calibration.

The major difficulty is the lack of observational data over the entire depth of oceans. Satellite radiometers have been monitoring global sea surface temperature (SST) for over 15 years, but the steric sea-level change depends on the temperature variations in the oceans at all depths (mainly in the mixing layer), which are only available from in situ observations, such as XBT and MBT. Due to the inadequate spatial and temporal coverage of historical in situ measurements, a global real-time three-dimensional (3-D) temperature field of the oceans is not available. However, if we focus on large-scale seasonal variations, the climatological temperature fields based on historical in situ measurement

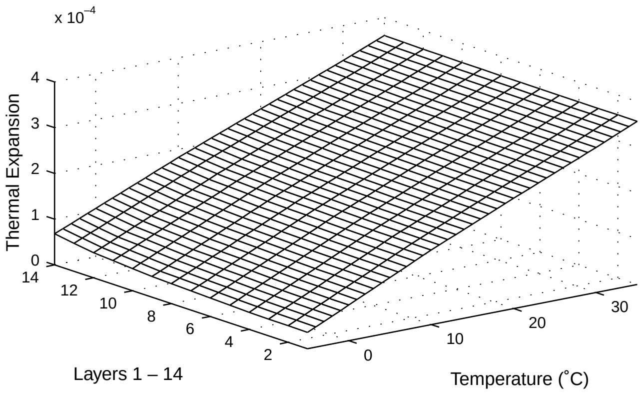

Fig. 2. The interpolated thermal expansion coefficients as a function of temperature (−5 to 35 °C) and the mean pressure of the top 14 layers of WOA94

will be valuable. The satellite SST data provide a means to improve the accuracy for the surface layer(s), especially in the southern oceans.

The steric contribution to global mean sea-level (MSL) change has been recently studied by Chen et al. (1998a) and Minster et al. (1999) using the climatological ocean temperature data in the NOAA World Ocean Atlas 1994 (WOA94) (Levitus and Boyer 1994). These two studies only discuss the global mean steric effect. In this study, we estimate steric sea-level change using both climatological data and satellite SST measurements at different large spatial scales, including basin, zonal, and global scales. The results are compared with corresponding sea-level changes from TOPEX/Poseidon satellite radar altimeter measurements. A 1° × 1° (latitude × longitude) monthly steric sea-level change data set, covering the same period as the TOPEX/Poseidon mission, is derived in this study. At the end, we discuss some error sources that may affect this computation.

2 Data and models

2.1 TOPEX/Poseidon observations

We use the TOPEX/Poseidon sea-level measurements of 10-day cycles 2 to 183 for the time period from October 1992 to July 1997. The media, instrument, and geophysical corrections applied include ionosphere delay, wet and dry troposphere delay, electromagnetic bias, tides, and the inverted barometer response. The TOPEX Geophysical Data Records (GDRs) orbits have been replaced with the new orbits computed with the JGM-3 gravity field model (Tapley et al. 1996), and the ocean tide model has been replaced with the UT/CSR 3.0 model (Eanes and Bettadapur 1995). The pole tide has been corrected. Sea-level anomalies relative to a mean sea-surface over 4 years are computed by interpolating the data to a fixed grid and then removing the mean sea-surface height (Chambers et al. 1997). The sea-surface anomalies are then averaged into a uniform 1° × 1° grid for each repeat cycle.

2.2 WOA94 climatology

The WOA94 objectively analyzed 1° × 1° monthly mean temperature fields represent 3-D seasonal ocean temperature variations (Levitus and Boyer 1994). These mean fields include 19 layers extending from the surface to 1000 m depth, and are derived from over 4.4 million historical in situ temperature profiles collected from various instrument groups, including XBT, MBT, CDT, and traditional bottle measurements. About 0.5 million are in the southern hemisphere and 0.1 million south of 40° latitude in the southern hemisphere. The WOA94 mean temperature fields are thought to represent seasonal variations in the northern hemisphere and tropical regions fairly well (Levitus and Boyer 1994; Chambers et al. 1997). The relatively sparse measurements in the southern hemisphere, especially south of

40°, implies that the actual spatial resolution is much worse than $1^\circ \times 1^\circ$ in the southern hemisphere. In addition to the monthly mean temperature fields, an objectively analyzed annual mean temperature field is also produced from all temperature profiles regardless of season or year.

2.3 Optimum interpolation SST data

The Optimum Interpolation SST (OISST) data are produced by combining satellite radiometer measurement, in situ observation, and the sea-ice model at the National Center for Environmental Prediction (NCEP) (Reynolds and Smith 1994). The monthly OISST fields are derived by a linear interpolation of the weekly optimum interpolation fields to daily fields, then the daily values are averaged over a month. The analysis uses in situ observation, satellite radiometer SST measurement, and the SST simulated by sea-ice cover. The in situ data are obtained from radio messages carried on the Global Telecommunication System, and the satellite measurements are from the operational data produced by the National Environmental Satellite, Data and Information Service (NESDIS). The spatial resolution of OISST fields is $1^\circ \times 1^\circ$. The OISST data we applied span the period from October 1992 to July 1997, the same period as that of the TOPEX/Poseidon data used here.

2.4 Improved 3-D ocean temperature field

Figure 3 shows comparisons of global mean SST deviations from annual mean field between the NCEP OISST data and the WOA94 climatological fields in March and September, respectively. In order to minimize the influence of interannual variations, the OISST mean fields are computed from all data in March or September. The SST differences between the OISST and WOA94 data in the two months are shown in the two bottom panels (e, f) of Fig. 3. In the tropical regions and subtropical regions in the northern hemisphere, WOA94 climatological data are in good agreement with the

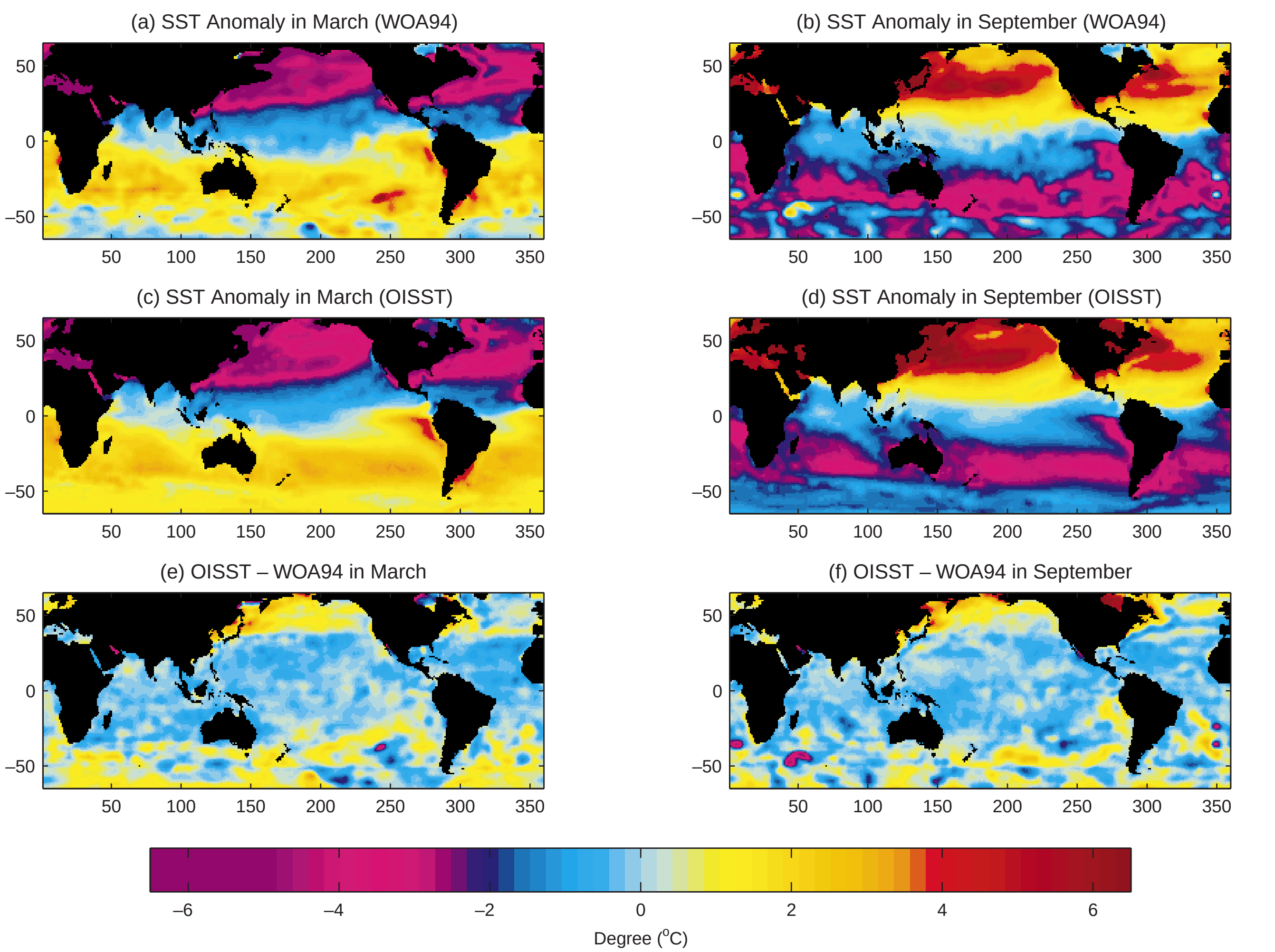

Fig. 3. **a, b** Climatological SST anomalies field in March and September with respect to the annual mean from the WOA94 objectively analyzed temperature; **c, d** the mean OISST anomalies in March and September, calculated using 4 years' OISST data (March 1993–February 1997); **e, f** SST differences between OISST and WOA94 SST (OISST–WOA94) in the two months

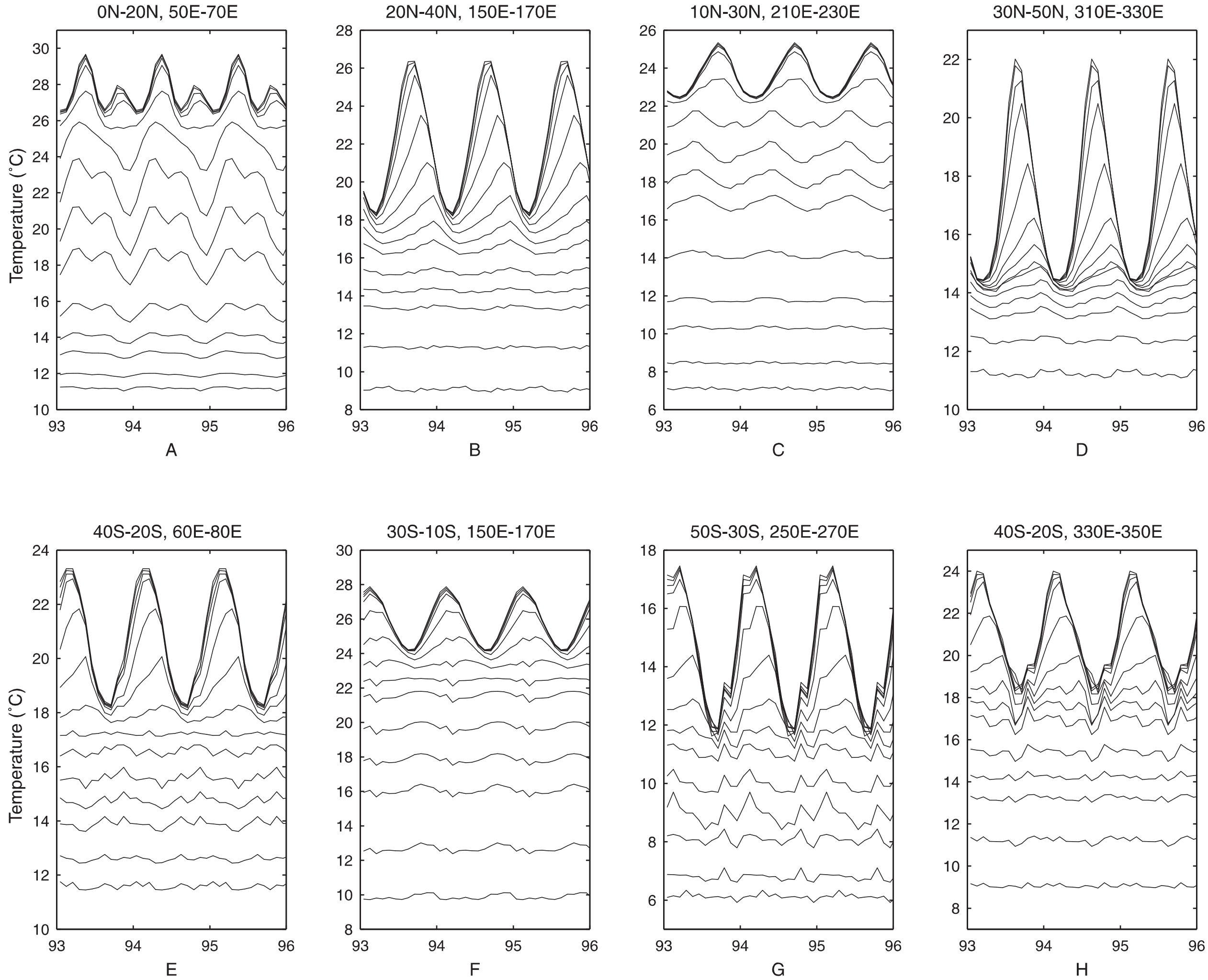

Fig. 4. The geographic distribution of the eight selected regions (A–H, corresponding to the temperature profiles shown in Fig. 5)

NCEP OISST data, while in the southern hemisphere and high-latitude regions in the northern hemisphere, discrepancies are obvious. The SST differences are less than ± 1 °C in most regions (Fig. 3e, f) and could be as large as ± 2–4 °C in the circum-Antarctic and high-latitude (northern hemisphere) regions.

The large discrepancies in the southern hemisphere and high-latitude regions are mainly due to the sparse historical in situ data samples in those regions and the different time spans used in computing the seasonal average. The WOA94 climatology is estimated from all available historical in situ temperature observations collected over a long period (over a century), while the OISST seasonal average is computed from only 6 years' measurements. The strong interannual and decadal SST variations will be a major error source to these discrepancies. The long-term SST change due to global warming will also contribute significantly to the differences. Part of the large positive discrepancies (OISST – WOA94) in high-latitude regions in the Pacific and Atlantic (see Fig. 3e, f) is probably due to long-term SST increase associated with global warming effects.

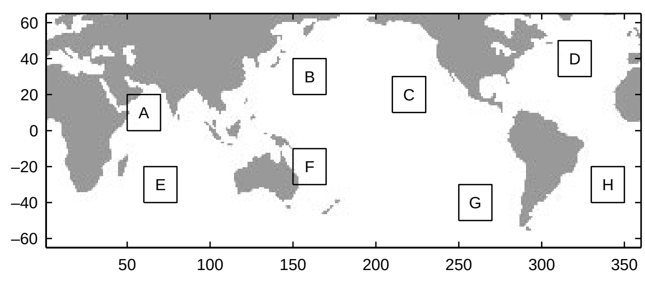

Fig. 5. Eight climatological predictions of vertical temperature profiles (A–H) in the northern and southern Indian, Pacific, and Atlantic Oceans. Each region is a $20° \times 20°$ (latitude by longitude) box

A natural approach to increasing the accuracy of the WOA94 climatology is to use the NCEP OISST data to improve the WOA94 mean temperature fields at the surface. Vertical temperature profiles will play a key role in determining how to effectively combine these two data sets. We chose several regions to study the mean vertical temperature profile using both the WOA94 cli-

Table 1. The standard depth definitions and the mean thickness of the top 14 layers of the WOA94 seasonal mean temperature fields

Layer	Depth (m)	Thickness (m)
1	0	5
2	10	10
3	20	10
4	30	15
5	50	22.5
6	75	25
7	100	25
8	125	25
9	150	37.5
10	200	50
11	250	50
12	300	75
13	400	100
14	500	100

matological temperature field and historical in situ temperature profiles. Figure 4 shows the geographical distribution of eight selected regions in the northern and southern Indian, Pacific, and Atlantic oceans. These eight regions are nearly randomly selected and the only consideration is to have an even coverage of the oceans. The temperature profiles of these eight regions are shown in Fig. 5A–H. In most regions, the mean temperatures of the top three layers (the top 0–25 m of the mixing layer), are very close to the surface temperature field. Based on this finding, and the study of other areas, we use the NCEP OISST data to replace the top three layers of our proxy 3-D ocean temperature field, and retain layers 4–14 (25–500 m depth) from the WOA94 climatology. The choice of a 25-m (i.e., the thickness of the top three layers of WOA94) replacement is conservative. We need to balance two kinds of error sources: one is from WOA94 due to poor sampling (especially in the southern hemisphere and high-latitude regions) and long-period average effects, and the other is from OISST due to vertical temperature variations. If we choose a deeper layer (e.g. 50 or 100 m), we will significantly increase errors in steric estimates by neglecting the temperature decrease in deeper layers.

3 Thermal expansion model

As we discussed above, in this study we focus on steric sea-level change due to thermal expansion. The thermal expansion coefficient C is defined as (Knauss 1979)

$$C = -\frac{1}{\rho}\frac{\partial \rho}{\partial T} \tag{1}$$

in which T is temperature; ρ is density of seawater; C is a function of temperature (T), pressure (P), and salinity

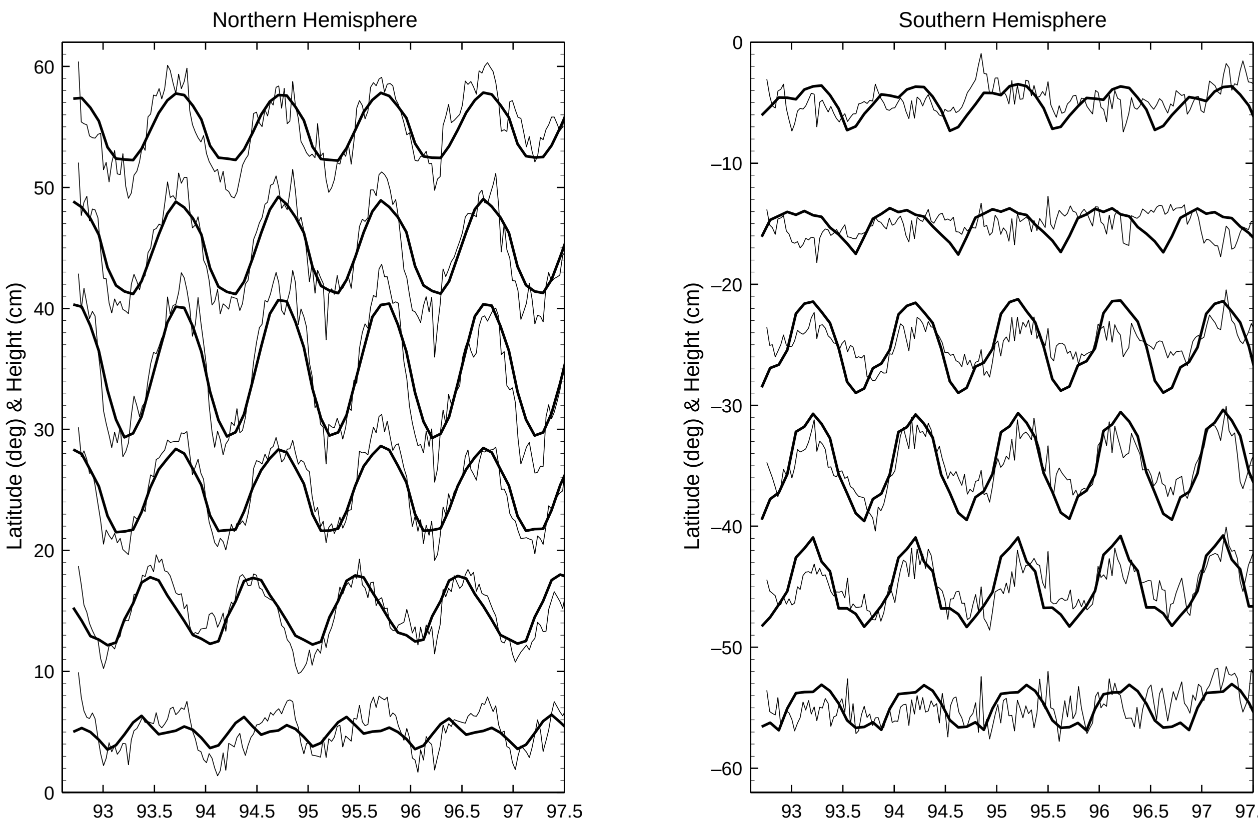

Fig. 6. The global zonal MSL variations in each 10° latitude band in **a** the northern hemisphere and **b** southern hemisphere. The *thin solid lines* represent the TOPEX/Poseidon observation, and the *thick solid lines* are the steric model prediction. The time series are vertically offset to corresponding latitude zone by adding the mean latitude in degree

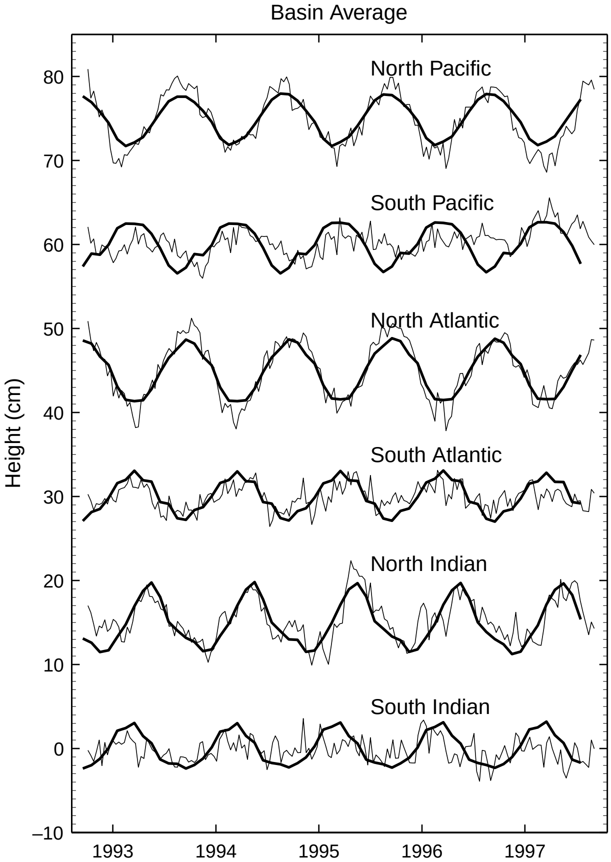

Fig. 7. The basin averages of MSL height in the northern and southern Pacific, Atlantic, and Indian Oceans. The *thin solid lines* represent the TOPEX/Poseidon observation, and the *thick solid lines* are the steric model predictions. The time series are offset vertically for a concise display

(S). Compared to temperature and pressure, the salinity effect on C is very small, and the salinity of the top few hundred meters of seawater is close to a constant (35‰). The experimental values of C as a function of T, P, and S given by Knauss (1979) are adopted in this model. We have interpolated the original C table into 1° intervals from −5 to 35 °C and to the mean pressure of each of the 14 layers of the proxy ocean temperature field described in Sect. 2.4. Figure 2 shows the interpolated thermal expansion coefficient matrix.

From the definition of thermal expansion coefficient C [Eq. (1)], it is quite easy to derive the sea-surface height change (ΔH) due to temperature variations (ΔT) under the assumption that there is no horizontal expansion. This assumption is accurate as long as the relative changes of ocean area with respect to the total ocean area due to sea-level variations are negligible. This is apparently the case in reality (even for enclosed and semi-enclosed seas). However, it will be questionable in coast regions. If the thickness of the seawater layer is H, mean temperature is T, and mean pressure is P, the steric sea-level change (ΔH) is

$$\Delta H = C(T,P) \cdot \Delta T \cdot H \tag{2}$$

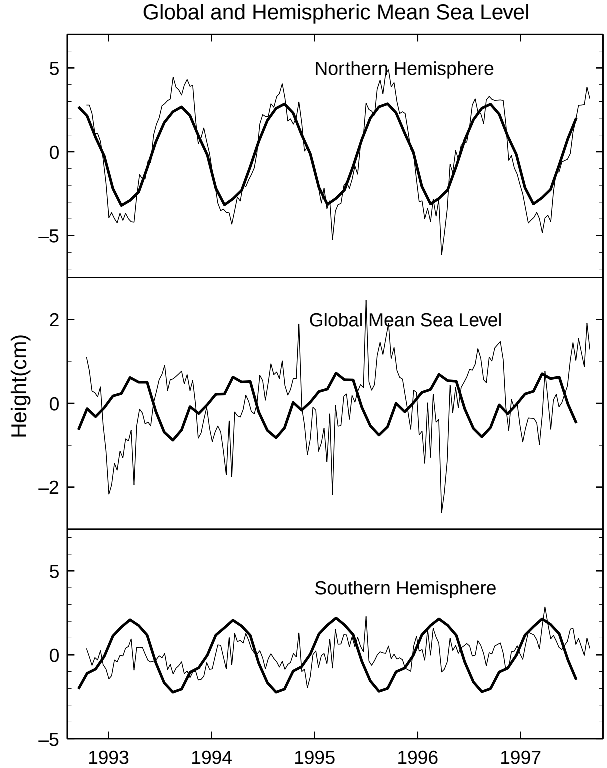

Fig. 8. The global and hemispheric MSL variations from TOPEX/Poseidon measurement and steric model estimation. The *thin solid lines* represent the TOPEX/Poseidon observation, and the *thick solid lines* are the steric model predictions

In order to derive the total steric sea-level change at a given grid point (latitude α, longitude λ) from the proxy real-time ocean temperature field, we simply perform the integration over the 14 layers. The temperature deviation $\Delta T(\alpha,\ \lambda,\ i,\ t)$ is calculated from the temperature variation relative to the annual mean (the 4-year mean for OISST and the climatological annual mean for WOA94) at each given point (α, λ), in each layer (i), at a given time (t).

$$\Delta H(\alpha,\lambda,t) = \sum_{i=1}^{14} C(T_i,P_i) \cdot \Delta T(\alpha,\lambda,i,t) \cdot H_i \tag{3}$$

Table 1 lists the thickness (H_i) of each of the 14 layers. One should be aware that if we neglect the salinity effect on steric sea-level change, this algorithm to compute steric sea-level change via thermal expansion coefficient is identical to the approach using density variation as applied in Minster et al. (1999).

4 Results and comparisons

4.1 TOPEX/Poseidon sea-level variation

Using the sea-level anomaly grids determined from TOPEX/Poseidon, we calculate the mean sea-level variations at different spatial scales. Figure 6 shows the zonal mean sea-level variations in each 10° latitude band

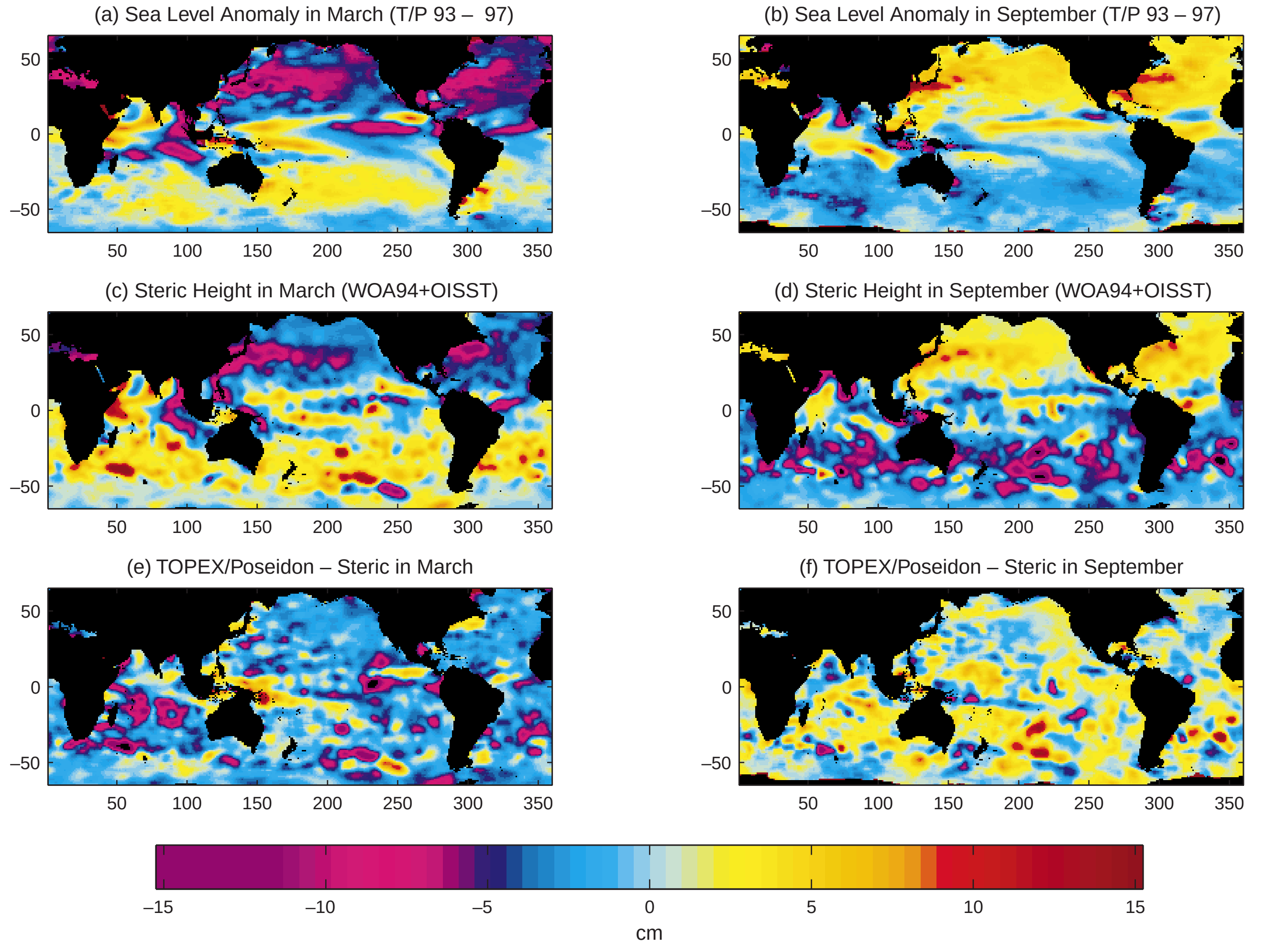

Fig. 9. **a**, **b** MSL anomalies determined by TOPEX/Poseidon in March and September; **c**, **d** mean steric sea-level variations determined by WOA94 and OISST data during the same time. The average is calculated using all observations in March regardless of years in both TOPEX/Poseidon data and the steric model grids; **e**, **f** sea-surface height differences between TOPEX/Poseidon observation and steric estimate in March and September, respectively

from 60°S to 60°N (i.e., 60°S–50°S, 50°S–40°S, . . . , 40°N–50°N, 50°N–60°N). An area weighting function (i.e., each grid point is weighted by cosine of the latitude of the grid) has been applied in calculating the mean sea-level change. Strong seasonal variations exist in most latitude regions, especially in the northern hemisphere. The dominant seasonal variation is found at the mid-latitudes near 30°N–50°N in the northern hemisphere with a mean seasonal variability of about 5–6 cm. The seasonal signal in the southern hemisphere is most notable at mid–high-latitude regions (20°S–50°S) with a mean annual amplitude near 2–3 cm. The maximum and minimum sea-surface heights in the northern hemisphere are in late summer/early fall (September) and late winter/early spring (March), respectively, and the phase is reversed in the southern hemisphere, with a maximum in March and a minimum in September. The seasonal variation almost disappears in circum-Antarctic and southern tropical regions.

Figure 7 shows the MSL variations for northern and southern Pacific, Indian, and Atlantic Oceans, and Fig. 8 shows the global MSL variation and the MSL variations in the northern and southern hemispheres. The seasonal variation is evident in different basins, especially the strong seasonal signals in the northern Pacific and Atlantic with a mean annual variability around 4–5 cm. The global MSL variation also shows a clear seasonal signal of nearly 1 cm amplitude, and has the same phase as the northern hemisphere (maximum in September and minimum in March).

In addition to the MSL time series, sea-level variation can be directly described by 2-D sea-level anomalies, especially for some regional seasonal patterns with small spatial scales. In order to minimize the interannual variability and high-frequency signals, we have calculated the monthly MSL anomaly field using all TOPEX/Poseidon cycles in a given month regardless of years. For example, the mean sea-level anomaly field in March is calculated by averaging a total of 15 cycles (18, 19, 20, 54, . . . ,) in March during the nearly five years from October 1992 to July 1997. The MSL anomalies in March and September are shown in Fig. 9a, b. The

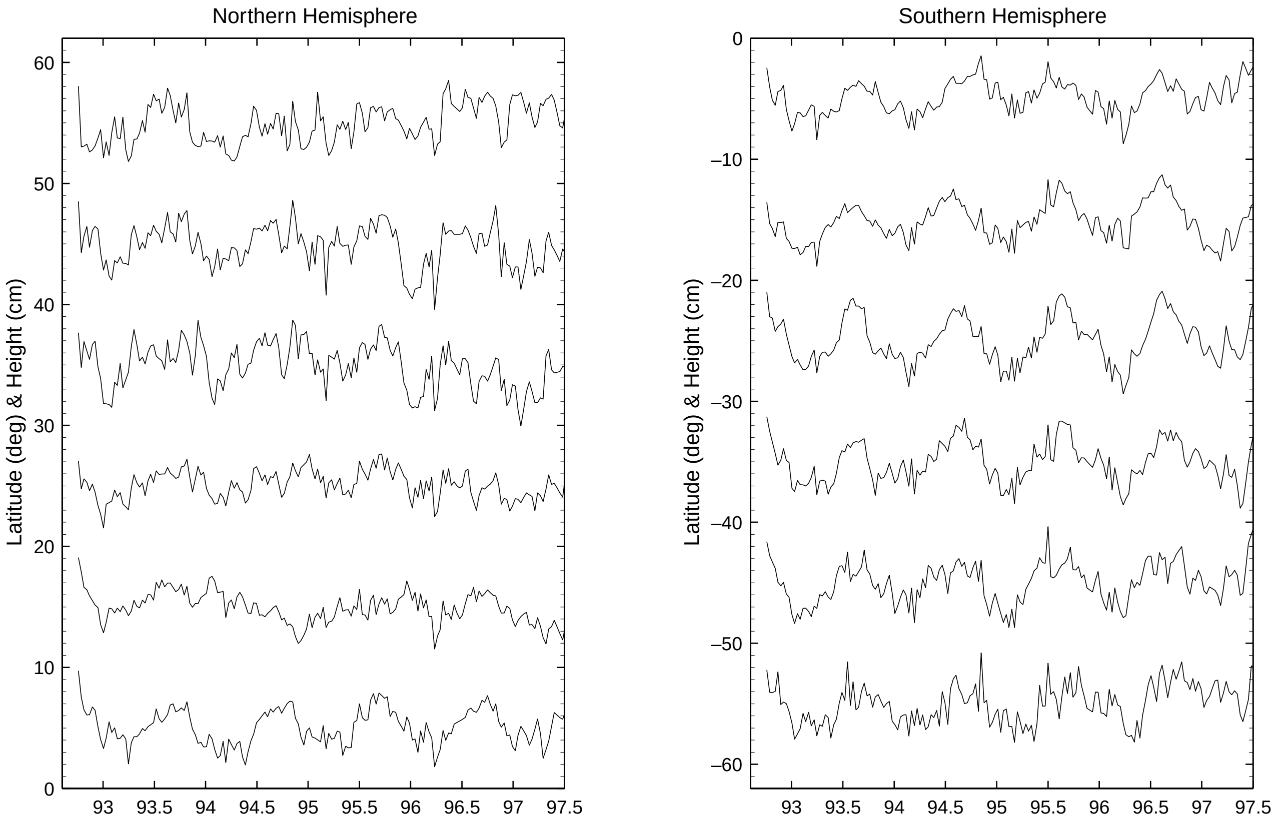

Fig. 10. MSL residuals in each 10° latitude zone (60°S–60°N) of the northern and southern hemisphere after steric effects are removed from TOPEX/Poseidon measurement

seasonal sea-level changes in the northern and southern hemispheres and the flipped phases are well represented in these two plots. Although in the MSL time series (see Fig. 6) seasonal signal in the tropical regions is not obvious, there are clearly some strong seasonal changes in these regions, especially the narrow zonal banded patterns in the tropical Pacific and Atlantic, and the less regular regional patterns in the Indian Ocean.

4.2 Steric sea-level change and comparisons

Monthly steric sea-surface height grids ($1° \times 1°$) are derived using the proxy real-time 3-D global temperature field and the thermal expansion model discussed above. We calculated the mean steric sea-surface heights for the same 10° latitude zonal regions as used in the TOPEX/Poseidon data, and plotted them together with the TOPEX/Poseidon observations in Fig. 6. The zonal mean steric sea-surface height variations show strong and dominant seasonal signals in both the northern and southern hemispheres, with a larger seasonal variability in the northern hemisphere (4–5 cm) than in the southern hemisphere (2–4 cm). The model-derived zonal mean steric sea-level variations are in very good agreement with the TOPEX/Poseidon observations, especially in the northern hemisphere, e.g., in the 30°N–40°N band, both TOPEX/Poseidon observation and steric estimate indicate a strong seasonal variability of $\pm$5–6 cm with the same phase (maximum heights in about September). Some larger discrepancies occur in the tropical and the high-latitude regions, i.e., the circum-Antarctic regions. The differences between TOPEX/Poseidon measurement and steric estimate are shown in Fig. 10.

The mean steric sea-level variations in the northern and southern Pacific, Atlantic, and Indian oceans are also calculated and compared with the TOPEX/Poseidon results in Fig. 7. Accordingly, the model-derived global steric sea-level variation is shown in Fig. 8, together with the estimates in the northern and southern hemispheres. Our model-derived steric sea-level variations also provide very good agreements with the TOPEX/Poseidon measurements at basin scales, especially in the northern Pacific, Atlantic, and Indian Oceans. However, in the southern hemisphere, the estimated steric heights are generally larger than the TOPEX/Poseidon observations, and the TOPEX/Poseidon observed global mean sea-level variation is out of phase with our steric model prediction (see Fig. 8). In the next section, we will discuss some possible causes for these discrepancies.

In order to look at some regional seasonal patterns we have seen in the TOPEX/Poseidon data, such as the zonal banded structures in the tropical Pacific and Atlantic Oceans, and the less regular seasonal patterns in the tropical Indian Ocean, we calculate the mean steric sea-surface heights in March and September (using the same average scheme applied in the TOPEX/Poseidon data), and show them in Fig. 9c, d. These regional seasonal patterns are well predicted by our steric model, especially the banded structure in the tropical Atlantic and the irregular seasonal patterns in the tropical Indian Ocean. The double-banded pattern in the tropical Pacific is also seen in the model estimate, but not as clearly as that in the TOPEX/Poseidon observations.

5 Discussion

This study confirms that thermal expansion is responsible for much of the large-scale seasonal variations observed by TOPEX/Poseidon altimeter, especially when averaging over zonal regions and basin scales (80–90% of the seasonal variability in the northern hemisphere; see Fig. 6). Some regional seasonal patterns in the tropical regions are also well predicted by this steric model. The results from this study indicate that historical in situ observational data and satellite SST data are valuable in determining seasonal steric sea-level height. If one could remove mass signals via independent techniques, e.g. using the future Gravity Recovery and Climate Experiment (GRACE) mission's observations, satellite altimeter measurements would lead to a better understanding of the steric sea-level and heat-storage change within the oceans. The latter bears a significant role in understanding global climate change.

The large discrepancies, especially those in the southern hemisphere, are mainly due to the sparse temperature measurements (Fig. 9e, f). Another major factor is that the TOPEX/Poseidon results are averaged over a 5-year period (with frequent El Niño events), and the steric estimates are primarily based on climatological data averaged over a much longer time period. The sea-level height changes from TOPEX/Poseidon measurement and steric estimate both indicate a considerably stronger seasonal signal in the northern hemisphere than those in the southern hemisphere. The relative weaker seasonal variability could be another reason for the poorer agreement in the southern hemisphere.

The phase difference in global MSL variations between TOPEX/Poseidon observation and steric model prediction (Fig. 8) implies a systematic difference between the two determinations. In addition to the error sources mentioned above, this may come from a variety of other sources. The water mass redistribution between the oceans, atmosphere, and continental water cycle (including snow/ice sheets) may have significant effects on TOPEX/Poseidon observed global MSL change, which are not part of and not included in the steric estimate. Recent studies (Chen et al. 1998a; Minster et al. 1999) have indicated that the seasonal variations of continental water storage change and atmospheric water vapor fluctuation together may introduce about 1 cm of changes in global MSL. Their estimated global MSL change based on hydrological models agrees very well (within a couple of mm in amplitude) with TOPEX/Poseidon observation if the steric effects are removed.

The systematic difference (shown in Figs. 8 and 10) with a mean seasonal variability at a 1–2 cm level could also be from the geophysical corrections applied in the TOPEX/Poseidon data, especially the IB correction and wet tropospheric delay correction. It is generally believed that the standard IB correction (i.e., using a constant reference pressure; Callahan 1993) applied in TOPEX/Poseidon data will introduce an incorrect seasonal signal (about 5 mm in amplitude) in global MSL determination (Raofi 1998). If we do not apply any IB correction (Chen et al. 1998a) or apply a modified IB model, e.g., using a non-constant reference pressure (Minster et al. 1999), the magnitude of seasonal global MSL change will be significantly reduced to several mm.

The seasonal steric sea-level change associated with temperature variation over the oceans is vital information required to separate mass variation from observed sea-level anomalies. The 1–2 cm residual signals (averaged over large spatial scales) are significant enough to be the vital driving forces for many geodynamical variations, e.g., the Earth rotation and gravitational field variation (Wilson 1995; Chen et al. 1998b) and geocenter motion (Chen et al. 1999), and provide observational constraints on the global water mass budget for the continental water cycle and atmospheric water vapor (Chen et al. 1998a; Minster et al. 1999). This research is an attempt to investigate this challenging problem using satellite SST measurement and historical in situ observation. The derived monthly $1^\circ \times 1^\circ$ grids of steric sea-level change will provide a chance to study the mass variation within the oceans using TOPEX/Poseidon (including its extended mission Jason-1) measurements, which play key roles in studying oceanic impacts on geodetic observations.

We have combined the NMC Optimum Interpolation SST data with the WOA94 objectively analyzed temperature fields to produce a proxy 3-D global ocean temperature model. This approach is based on the fact that seasonal signals are dominant in the top few hundred meters, and the mean temperature in the first few layers are quite close to the sea-surface temperature for the areas we examined (see Fig. 5). This combination is applied to improve the accuracy of WOA94 mean temperature fields, especially in the southern hemisphere. Because the physical process of downward heat flux transfer from the surface is complicated, one cannot simply infer the vertical temperature profile from surface temperature data without completely understanding the process, even if we know the mixing layer depth. On the other hand, the deep layers (4–14 in this model) provide significant contributions to the total steric sea-surface height due to the comparable variability of seasonal temperature variation (see Fig. 5), so historical in situ temperature data play an important and unique role in studying steric sea-level change.

Quantitative assessment of the improvement by using OISST data is difficult to estimate at the moment. A better agreement with TOPEX/Poseidon measurement is not an essential indicator of the improvement. This is mainly because non-steric sea-level variations (or mass-change-associated sea-level variations) obviously exist within the oceans, and are clearly demonstrated in some recent studies (Chen et al. 1998a, b, 1999; Minster et al. 1999). A reliable assessment of the improvement can only be estimated if we are able to remove the non-steric signals via independent techniques, e.g., using the future GRACE observations. However, we are confident of the improvement because the OISST temperature field is widely considered a better description of the surface temperature variation than the WOA94 climatology, especially in the southern hemisphere (Reynolds and Smith 1994).

This approach has many limitations and is only useful in the first order of approximation. Because the temperature data of only the top 25 m are mainly from real-time observations (satellite SST), which is a small portion of the total mixing layer, this study is not able to provide reliable estimates of the interannual steric sea-level change. However, we do see some interannual steric sea-level change signals in specific regions, e.g. the eastern and western tropical Pacific, which are correlated with TOPEX/Poseidon measurements, although the amplitudes are much smaller (10–20% of what we get from TOPEX/Poseidon determination in the eastern Pacific, for example). In some specific regions (e.g., region D in Figs. 4, 5), the temperature differences between the surface and the 25-m depth are as large as 1 °C. So, this direct combination will overestimate the steric effects, in some senses. The selection of different temporal references (4-year mean or long-term mean) and reference depths will also affect the steric estimates. As more in situ observational data (like XBT, MBT) and satellite SST data become available, we can expect a more reliable 3-D ocean temperature field, which will lead to a further understanding of steric sea-level change at various temporal and spatial scales over the oceans. A challenging task is how to dynamically integrate historical in situ measurements, remote sensing data, and recent XBT observations into a physically coherent 3-D ocean temperature model via objective analysis or data assimilation.

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