ABSTRACT. The purpose of this paper is to estimate long-term and short period oscillations in global sea level changes using sea level anomalies data computed from TOPEX/Poseidon altimetry. The long-term global sea level rise computed by the robust method is of the order of 1.5 mm/year. The highest sea level rise of the order of 10-20 mm/year can be noticed in the east-central Indian and west-central Pacific Oceans. Short period oscillations with periods from about 20 to 200 days of sea level anomalies data computed from TOPEX/Poseidon altimetry are detected using the Fourier transform band pass filter. The mean amplitude spectra of these data for the whole ocean, northern and southern hemispheres show peaks corresponding to the semiannual, 120, 90, 62, 37 and 30-day oscillations. The maps of the mean amplitudes of these oscillations in sea level anomalies data as well as the same data with no inverse barometer correction are presented. The amplitudes of these oscillations are variable in time and the maps of time variable amplitudes of the semiannual oscillation are presented. Global maps of the mean amplitudes of 37 and 30-day oscillations reveal a pattern of regular trains in the equatorial central Pacific Ocean with the amplitudes of the order of 1 cm caused by the Legeckis waves also known as tropical instability waves.

1. INTRODUCTION

The topography of the sea-surface relative to Earth's geoid is a basic measurement carried now by satellite altimetry which provides the global, near-instantaneous measurement of the ocean's large-scale circulation and its variability. Altimetry data allows to study global ocean circulation and the ocean's influence on weather and climate. TOPEX/Poseidon (T/P) is a satellite mission that uses radar altimetry to make precise measurements of sea level with the primary goal of studying the global ocean circulation. T/P satellite measures global sea level with unparalleled accuracy of the order of 2 cm (Fu et al. 1994, Cheney et al. 1994) and such accuracy allows to study variations in global sea level heights. Such high accuracy of global sea level is possible due to precise orbit of the T/P (Tapley et al. 1994).

The ocean sea level shows periodic variations ranging from days to centuries. Global-average sea level is believed to have risen by between 10-20 cm during the past century due to global warming from the greenhouse effect (EPA 2001). Rising sea levels are largely a consequence of the thermal expansion of the ocean, melting of low latitude glaciers in Alps and Rocky Mountains, and many
other factors reviewed every few years by the Intergovernmental Panel on Climate Change (IPCC) (PSMSL 2001). Climate models used to study greenhouse effect predict an increase of global mean sea level by 30-50 cm in this century (Church et al. 1991). The rate of global sea level rise estimated from tide gauges measurements is of the order of 1.5 to 2 mm/year (Trupin and Wahr 1990, Douglas 1991). The global mean sea level rise estimated from T/P altimeter data (cycles from 9 to 94) is of the order of 5.8 mm/year (Nerem 1995). The global sea level rise estimated recently by Shum et al. (2001) from all available altimetric and tide gauge measurements is of the order of 1.5 mm/year. The annual and semi-annual oscillations were found by Nerem et al. (1994), Knudsen (1994), Wunsch and Stammer (1995), Chang Kou Tai (1996), Kosek and Kolaczek (1999) and their amplitudes depend on geographic regions. The annual and semiannual oscillations can reach the values of 10-15 cm and 5-7 cm, respectively in some geographic regions of the northern hemisphere (Nerem et al. 1994, Knudsen 1994). The periodograms of global mean sea level variations shows oscillations with periods of about 120, 90, 60, 50, 40, and 30 days (Nerem 1995). Amplitudes of these short period variations are of the order of 1-3 cm (Chao and Fu 1995). Short period variations with periods of 182, 120, 60 and 40 days in sea level anomalies (SLA) data with the amplitudes of the order of 1–7 cm were found also by Kosek and Kolaczek (1999). Their amplitudes vary in time as it was reported by Kosek (1999a,b).

The main purposes of this paper are to estimate:
- the rate of global sea level change for the whole ocean and as a function of geographic location,
- the periods and amplitudes of short period oscillations ranging from 20 to 200 days for the whole ocean and as a function of geographic location,
- amplitude variations of the semiannual oscillation as a function of geographic location and their relations with El Niño event in 1997/98.

2. SEA LEVEL ANOMALY DATA

Altimeter analyses from the T/P altimeter mission are provided in terms of sea level deviations and anomalies. The SLA represents the difference between the best estimate of the sea surface height and a mean sea surface (Chambers et al. 1998). The mean surface topography is computed as heights of 4-year average sea surface height relative to the JGM-3 geoid (Tapley 1996). The sea surface height was corrected for atmospheric effects (ionosphere, wet and dry troposphere), effects due to surface conditions (electromagnetic bias), and other contributions (ocean tides, pole tide, and inverse barometer).

The SLA data were computed using a precise orbit of T/P computed in the University of Texas at Austin, Center for Space Research (UT/CSR), USA from satellite tracking data as described by Chambers et al. (1998) and Cheney et al. (1994). The altimeter measurements are reduced to sea surface topography heights using a precise orbit computed at UT/CSR using the available Satellite Laser Ranging (SLR) and Doppler Orbitography and Radiopositioning Integrated by Satellite (DORIS) tracking data. Precise orbits were computed with the Joint Gravity Model (JGM-3), the latest force models, and ocean tide correction based on the UT/CSR 3.0 model. Data are available along the satellite track or in gridded form. The gridded form data obtained after interpolating 1-second T/P altimetric measurements into 1°×1° grids are described detaily by CSR (2001). This data base contains gridded sea level anomaly heights called ‘cycles’ according to the T/P repeat cycle number from −65° to 65° latitude and from 0° to 359° longitude and their sampling interval is equal to $\Delta t = 9.9140625$ days. The estimated accuracy of SLA time series for 1°×1° grids is of the order of 3 to 4 cm over the entire ocean.
The cycles from 10 (Dec. 21\textsuperscript{th} 1992) to 309 (Feb. 2\textsuperscript{nd} 2001) (CSR 2001) were used for computation of the secular sea level change and amplitudes of short period oscillations. The missing cycles with the numbers 118, 174, 278, 289, 299 and 307 were interpolated between cycles by linear interpolation. Additionally, SLA data set (cycles from 10 (Dec. 21\textsuperscript{th} 1992) to 285 (Jun. 9\textsuperscript{th} 2000)) with no inverse barometer correction (NOIB) were used (CSR 2001). The missing NOIB cycles with numbers 118, 174, 197, 216, 224 and 278 were interpolated between cycles. Cycle 118 is missing because problems with the satellite caused by the altimeter shut down. Cycle 174 is missing because of a combination of problems with the Poseidon altimeter, the microwave radiometer, and the DORIS tracking system. Cycles 278, 289, 299 and 307 and NOIB cycles 197, 216 and 224 are missing due to some problems with Poseidon data that are usually noisier than TOPEX data.

To estimate the mean sea level rise for the whole ocean the monthly SLA cycles from 1 (January 1993) to 94 (November 2000) in $1\degree \times 1\degree$ grids were used (CSR 2001). In these data all standard corrections were applied, including the inverse barometer model.

### 3. VARIATIONS IN MEAN SEA LEVEL

The 10-day and monthly SLA data from $1\degree \times 1\degree$ geographic region with the latitude $\varphi$ and longitude $\Lambda$ were used to compute the mean sea level change using robust straight-line estimation (Priestley 1981). The maps of the mean sea level change for 10-day and monthly SLA data as a function of geographic latitude $\varphi$ and longitude $\Lambda$ are very similar (Fig. 1). The mean sea level change computed from these data shows about 10 mm/year decrease of sea level in the east-central Pacific Ocean and in the north-west and central-west Indian Ocean. The highest increase in sea level of the order of about 10-20 mm/year can be noticed in the west-equatorial Pacific Ocean and east Indian Ocean near the coasts of Indonesia, Philippines and New Guinea. Secular sea level change estimated from T/P altimetric data (cycles 9-94) shows the highest increase of sea level of the order of 80 mm/year in the central equatorial Indian Ocean and 40-60 mm/year in the western equatorial Pacific Ocean (Nerem 1995). Changes in ocean level due to climate change can be greater in some places than others because the ocean circulation will adapt to accommodate the new climate regime (PSMSL 2001).

![Fig. 1. Linear trend computed by the robust method from the 10-day (left) and monthly (right) SLA data as a function of geographic position.](image)

The mean sea level computed as the average value of 10-day and monthly SLA data of the whole ocean region shows annual variations (Fig. 2). The straight-line fits to these mean sea levels computed by the robust method (Priestley 1981) show that the mean rise rates are of the order of 1.5 mm/year for 10-day SLA data and 2.0 mm/year for monthly SLA data (Fig. 2). The difference in the mean sea level rise computed from the 10-day and monthly SLA data could be caused by the fact that the monthly SLA data are shorter than the 10-day data and during the last few years after
1999 the average value of the 10-day SLA data does not show any rise of the mean sea level. The highest increase in ocean sea level can be noticed during the last El Niño 1997/1998. Recent analyses by U.S. Geological Survey (USGS) scientists of nearly 100 years of sea-level records collected near the Golden Gate Bridge found that these abnormally high sea levels were the direct result of that year's El Niño atmospheric phenomenon (USGS 2001).

Fig. 2. Mean sea level rise rates computed by the robust method from 10-day (left) and monthly (right) SLA data.

4. THE FOURIER TRANSFORM BAND PASS FILTER

The 10-day SLA time series from 1° × 1° geographic region with the latitude \( \varphi \) and longitude \( \Lambda \) from which the linear trend computed by the robust method (Priestley 1981) was removed were filtered by the Butterworth high pass filter (HPF) (Otnes and Enochson 1972) with the cut-off period of 270 days to eliminate the annual and residual longer period variations. In order to investigate short period oscillations of SLA data the Fourier transform band pass filter (FTBPF) was applied (Popiński and Kosek 1995). The output of this filter is determined by the following formula:

\[
    u_{\varphi, \Lambda}(t, T) = FFT^{-1}\left[FFT[x_{\varphi, \Lambda}(t)]A(P, T]\right]
\]

where \( u_{\varphi, \Lambda}(t, T) \) is an oscillation with central period \( T \) computed by the FTBPF, \( x_{\varphi, \Lambda}(t) \) are the SLA time series with linear trend removed and filtered by the Butterworth HPF with the 270-day cut-off period, \( FFT \) is the Singleton (1969) Fast Fourier Transform operator and

\[
    A(P, T) = \begin{cases} 
        1 - \left( \frac{\Delta t/P - \Delta t/T}{\lambda} \right)^2 & \text{if } |\Delta t/P - \Delta t/T| \leq \lambda \\
        0 & \text{if } |\Delta t/P - \Delta t/T| > \lambda 
    \end{cases}
\]

is the parabolic transfer function, \( \lambda > \pi/N \) is the transfer function bandwidth. In this paper the value of \( \lambda = 0.015 \) was adopted due to a good resolution of detected oscillations.
5. MEAN AMPLITUDE SPECTRA

The FTBPF spectrum of 10-day SLA data from \(1^\circ \times 1^\circ\) geographic region with the latitude \(\varphi\) and longitude \(\Lambda\) is defined by the following formula (Kosek 1995, Kosek et al. 1998, Popiński and Kosek 1995, Kosek 1999a,b):

\[
\hat{S}_{\varphi,\Lambda}(T) = \frac{1}{N-2m} \sum_{k=m+1}^{N-m} \left| \hat{u}_{\varphi,\Lambda}(k,T) \right|^2 ,
\]

(3)

where: \(N\) is the number of cycles, \(m = 20\) is the number of points that must be dropped at the beginning/end of data due to filter errors and \(T\) is an oscillation period.

The mean spectrum of SLA data of the rectangular ocean area with the minimum and maximum latitudes and longitudes \(<\varphi_p,\varphi_k> \times <\Lambda_p,\Lambda_k>\) is computed by the following formula:

\[
\hat{S}_{\text{mean}}(T) = \frac{1}{K} \sum_{\Lambda=\Lambda_p}^{\Lambda_k} \sum_{\varphi=\varphi_p}^{\varphi_k} \hat{S}_{\varphi,\Lambda}(T)
\]

(4)

where \(K\) is the number of SLA time series in the rectangular ocean/land area.

The square root of the spectra (eqs. 3 and 4) multiplied by \(\sqrt{2}\) represent the amplitude spectra in which their values are proportional to the mean amplitudes of oscillations (Kosek 1995).

The mean amplitude spectra for the whole ocean \(<-65^\circ,65^\circ> \times <0^\circ,359^\circ>\), the northern \(<0^\circ,65^\circ> \times <0^\circ,359^\circ>\), and southern \(<-65^\circ,0^\circ> \times <0^\circ,359^\circ>\) hemispheres as a square root of the mean spectra \(\hat{S}_{\text{mean}}(T)\) (eq. 4) are shown in Figure 3. The number of SLA time series \(K\) for the whole ocean, the northern and southern hemispheres are equal to 33296, 13088 and 20208, respectively. The number of NOIB SLA time series \(K\) for the whole ocean, the northern and southern hemispheres are equal to 32972, 13007 and 19965, respectively. The mean amplitude spectra of SLA data show peaks corresponding to oscillations with periods of 183, 120, 90, 62, 37 and 30 days and their mean amplitudes decrease with an oscillation period. The mean amplitudes of oscillations in SLA data are smaller than the mean amplitudes of oscillations in NOIB SLA data. The mean amplitude spectra for the northern hemisphere have greater values than for the southern one. The SLA data are smoother than NOIB SLA data for oscillations with periods less than about 50 days, because their mean amplitude spectra are smoother in this shorter period band. It means that inclusion of the inverse barometric correction into SLA data diminish the mean amplitudes of all short period oscillations as well as smoothes these data.

**Fig. 3.** The mean amplitude spectra of the SLA (left) and NOIB SLA (right) data for the whole ocean (heavy line), northern (N – dashed line) and southern (S – dotted line) hemispheres.
The mean spectra as a function of geographic longitude and latitude are defined by the following formulae:

\[
\hat{S}_\Lambda(T) = \frac{1}{K(\Lambda)} \sum_{\theta=\varphi}^\varphi \hat{S}_{\varphi,\Lambda}(T), \quad \hat{S}_\varphi(T) = \frac{1}{K(\varphi)} \sum_{\Lambda=\Lambda'}^\Lambda \hat{S}_{\varphi,\Lambda}(T),
\]

respectively, where \( K(\Lambda) \) and \( K(\varphi) \) denote the number of SLA time series along the longitudes \( \Lambda \) and latitudes \( \varphi \), respectively.

The mean amplitude spectrum in latitude, as square root of the mean spectrum \( \hat{S}_\varphi(T) \) (eq. 5) shows that the semiannual oscillation is energetic mostly in the equatorial regions and its highest average value is of the order of 2.5 cm (Fig. 4). The semiannual oscillation is also energetic at mid-latitudes of the northern hemisphere as well as in both hemispheres at high latitudes greater than 60°. The oscillations with periods of 120, 90 and 60 days are less energetic than the semiannual one but they are present in the same latitude regions as the semiannual oscillation.

The mean amplitude spectrum in longitude, as square root of the mean spectrum \( \hat{S}_\Lambda(T) \) (eq. 5) shows that the semiannual oscillation is the most energetic at the longitudes from 20° to 60° and from 90° to 110° (Fig. 4). Usually, the oscillations with periods of 120, 90 are energetic at the same longitudes as the semiannual oscillation except the 62-day oscillation which is mostly energetic at longitudes 120° and 290°. The smallest amplitudes of the semiannual and other shorter period oscillations can be noticed at longitudes from 200° to 240° and from 320° to 350° that correspond to the central Pacific and central Atlantic Oceans, respectively. The presence of the 120, 90, 62-day oscillations in the same latitude and longitude regions as the semiannual oscillation may suggest that their character may be also seasonal.

**Fig. 4.** The mean amplitude spectra of the data as a function of geographic latitude (left) and longitude (right).

### 6. GLOBAL AMPLITUDE SPECTRA

The maps of the mean amplitudes of oscillations in SLA data with the central periods 183, 120, 90, 62, 37 and 30 days as a square root of the spectrum \( \hat{S}_{\varphi,\Lambda}(T) \) (eq. 3) are shown in Figure 5. The geographic locations of the highest amplitudes of these oscillations together with average amplitudes in these regions are given in Table 1. Usually, the mean amplitudes decrease with an oscillation period. The geographic regions of the highest amplitudes are almost common for all detected oscillations.
The map of the mean amplitudes of the semiannual oscillation is very similar to those computed by Nerem et al. (1994) as well as Kosek and Kołaczek (1999). The maps of the mean amplitudes of the 120 and 62-day oscillations are very similar to the maps of the mean amplitudes of 120 and 60-day oscillations computed previously from smaller number of T/P cycles by Kosek and Kołaczek (1999). The big amplitudes of the 62-day oscillation in some coastal regions of Indonesia, Malaysia and Philippines, the south-west coasts of Japan, and south-east coasts of Argentina are caused by the mismodelling of the \( M_2 \) and \( S_2 \) tide wave amplitudes since 62-day oscillation is an alias oscillation generated by measuring these tidal waves with a T/P sampling frequency equal to 9.9156 cycles/day (Wagner et al. 1994, Katz et al. 1995).

The maps of the mean amplitudes of oscillations with the central periods 183, 120, 90, 62, 37 and 30 days computed by the FTBPF (eq. 3) in NOIB SLA data are shown in Figure 6. This analysis shows that inclusion of the inverse barometer correction in SLA data diminishes the amplitudes of all short period oscillations especially at higher latitudes in both hemispheres.

6.1. Tropical Instability Waves

More than 20 years ago, Richard Legeckis (1977) used NOAA satellite observations to demonstrate that the westward slowly flowing South Equatorial Current moves in a wave-like pattern in the eastern equatorial Pacific. The satellite observations of equatorial Legeckis waves were confirmed in theoretical studies of ocean circulation by Cox (1980) who concluded that these waves were due to both baroclinic and barotropic instability. These waves are believed to be generated by instabilities of the equatorial ocean currents. During La Niña the equatorial waters cool down in response to the atmospheric winds and the interaction of the equatorial currents produces a phenomena known as the equatorial long waves or tropical instability waves (TIWs). Both satellite observations and ocean circulation models have shown that TIWs, also known as Legeckis waves, are westward propagating disturbances with wavelengths of 1000–2000 km and periods of 20-40 days as seen in satellite measurements of sea surface temperature (Allen et al. 1995, Lawrence et al. 1998, Liu et al. 2000, Hashizume et al. 2001).

TIWs have been investigated during Tropical Instability Wave Experiment (TIWE) in 1990-91 (Qiao and Weiberg 1995) using Tropical Atmosphere-Ocean (TAO) Array of moored ocean buoys in the tropical Pacific. TAO Array consists of nearly 70 moored buoys spanning the equatorial Pacific and measures oceanographic and surface meteorological variables for investigations of seasonal-to-interannual climate variations originating in the tropics, that are mostly related to El Niño/Southern Oscillation (ENSO) events.

There are some suggestions that ENSO can be interpreted in terms of the interaction of interannual Kelvin and Rossby waves in the Pacific ocean. A series of numerical experiments have been set up to examine the relationship between the intraseasonal Kelvin waves and TIWs. It is possible that the arrival of the oceanic Kelvin waves modifies the meridional current structures as to generate TIWs (Benestad 1997).

Chelton et al. (2000) examined the evolution of TIWs and found that TIWs in the Pacific are seen on both sides of the equator with about 50% larger amplitudes in the north than in the south. In the Atlantic, however, the TIWs are seen only north of the equator. The maps of the mean amplitudes of 37 and 30-day oscillations computed by the FTBPF in SLA and NOIB SLA data (Figs. 5, 6) show a pattern of regular trains with amplitudes of about 1 cm only north of the equator in central-east Pacific. This pattern of regular trains, caused by TIWs seems to be more regular for the 30-day oscillation than for the 37-day one.
Table 1. The geographic location of the oscillations in SLA data shown in Figure 5 and their average amplitudes.

<table>
<thead>
<tr>
<th>Oscillation period (days)</th>
<th>Geographic location (amplitude in cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>183</td>
<td>East and west Indian Ocean (6-7), Sea of Okhotsk (7-9), Baltic Sea (3-6), Red Sea (3-6), east-equatorial Atlantic Ocean – Gulf of Guinea (4-5), Gulf of Aden (3-5) Arabian Sea (3-4), east and west equatorial Pacific Ocean (2-4), Black Sea (2-4), Mediterranean Sea (2-3).</td>
</tr>
<tr>
<td>120</td>
<td>Baltic Sea (3-4), North Sea (2-4), Black Sea (2-3), Okhotsk Sea (2-4), Indian Ocean (1-2), west Pacific Ocean (2-3), west north Pacific Ocean (1-3), Argentinean Basin (1-3).</td>
</tr>
<tr>
<td>90</td>
<td>Sea of Okhotsk (3-6), Baltic Sea (2-4), central and east Indian Ocean (1-3), Red Sea (2-3), North Sea (2), east coast of Africa (1-2), Yellow Sea and East China Sea (1-2), Argentinean Basin (2-3), Australian-Antarctic Basin (1-2), Pacific-Antarctic Basin (1), North American Basin (1-2), Gulf of St. Lawrence (1-2), south-east coasts of Australia (1-2), South coasts of Africa (1-2).</td>
</tr>
<tr>
<td>62</td>
<td>Coasts of Indonesia, Malaysia and Philippines (4-9), Yellow Sea and East China Sea (4-9), Sulu, Sulawesi and Molucca Seas (4-9), Java Sea (2-4), Andaman Sea (2-5), Argentinean Basin (2-8), Gulf of Alaska (2-3), Bay of Funday and Gulf of St. Lawrence (2-8), Californian Bay (2-6), Red Sea (3-4), Baltic Sea (3-4), North Sea (2-3), Persian Gulf (2-3), Mozambique Channel (2), north-east coasts of Australia (1-2).</td>
</tr>
<tr>
<td>37</td>
<td>Baltic Sea (1-3), North Sea (2), Black Sea (1-2), Caspian Sea (1-2), Red Sea (2), east Mediterranean Sea (1), Gulf of St. Lawrence (2), Hudson Bay (2), Legeckis waves in central equatorial Pacific Ocean (1-2), Pacific-Antarctic Basin (1-1.5), Australian-Antarctic Basin (1-1.5), Argentinean Basin (0.5-2).</td>
</tr>
<tr>
<td>30</td>
<td>North Sea (2), Baltic Sea (2), Red Sea (2), Okhotsk Sea (2), Hudson Bay (1-2), Gulf of St. Lawrence and Bay of Funday (1-2), Argentinean Basin (0.5-2), Pacific-Antarctic Basin (1), Australian-Antarctic Basin (1), Yellow Sea and East China Sea (1), Legeckis waves in central equatorial Pacific Ocean (1), Gulf of Carpenteria (1-2).</td>
</tr>
</tbody>
</table>

7. GLOBAL TIME VARIABLE AMPLITUDE SPECTRA OF THE SEMI-ANNUAL OSCILLATION

The time variable FTBPF spectrum as a function of time $t$ and oscillation period $T$ is defined by the following formula (Kosek 1995, 1999a,b):

$$\hat{S}_{\phi,A}(t,T) = \frac{1}{M(T)} \sum_{k=-M(T)/2}^{M(T)/2} \left| \hat{u}_{\phi,A}(t + k, T) \right|^2$$

(7)

where $M(T) = 2T / \Delta t$.

The square root of the time variable spectrum multiplied by $\sqrt{2}$ represent the instantaneous amplitude of oscillations with central period $T$. The time variable amplitude spectra of the semiannual oscillation in SLA data are shown in spring of each year from 1993 to 2000 in Figure 7. The semiannual oscillation is the most energetic in the equatorial Indian and Pacific Oceans, some of the Europeans seas and the Okhotsk Sea. It can be noticed that the last El Niño event in
1997/98 was preceded by increase of the amplitude of the semiannual oscillation in the east and west Indian Ocean in spring 1996. During the El Niño 1997/98 an increase of the amplitude of the semiannual oscillation can be noticed in the east equatorial Pacific Ocean together with simultaneous decrease of its amplitude in the Indian Ocean. Beginning of the last El Niño 1997/98 is also associated with an increase of the amplitude of the semiannual oscillation in the Baltic, Red, Black, Caspian, east Mediterranean Seas and Persian Gulf (Fig. 7). The amplitude of the semiannual oscillation diminish in all these mentioned regions in 1999 during La Niña. Very high amplitude of the order of 10 cm of the semiannual oscillation can be noticed in the Baltic Sea in spring of 1993.

Fig. 7. The maps of time variable amplitudes of the semiannual oscillation filtered by the FTBPF (eq. 3) in SLA data.
8. CONCLUSIONS

The global sea level rise and amplitudes of short period oscillations with periods from 20 to 200 days were estimated from SLA data of T/P altimetry. The mean sea level rise for the whole ocean estimated from SLA data is of the order of 1.5 mm/year and it is similar to this estimated by Shum et al. (2001) from all available altimetric and tide gauges measurements. The highest sea level rise of the order of 10-20 mm/year can be noticed in the east-central Indian and west-central Pacific Oceans.

Mean amplitudes of 183, 120, 62, 37 and 30-day oscillations of SLA data computed by the FTBPF depend on geographic position. Usually, the mean amplitudes of these oscillations decrease with an oscillation period. The geographic regions of the highest amplitudes are almost common for all detected short period oscillations and similar to the highest amplitudes of the semiannual oscillation, what suggests their seasonal character. The big amplitudes of the 62-day oscillation in same costal regions of Indonesia, Malaysia an Philippines, Yellow Sea and East China Sea, southeast coasts of Argentina and possibly other costal regions are caused by the missmodelling of the $M_2$ and $S_2$ tide wave amplitudes.

The mean amplitudes of 37 and 30-day oscillations of SLA and NOIB SLA data show a pattern of regular trains with amplitudes of about 1 cm in central-east Pacific Ocean that are caused by TIWs. This pattern of regular trains seems to be more regular for the 30-day oscillation than for the 37-day one. Thus, application of the FTBPF to the SLA data for the first time enabled detection of the TIWs from space.

The El Niño in 1997/98 was preceded by increase of the amplitude of the semiannual oscillation in the east and west Indian Ocean in spring 1996 as well as increase of the amplitudes of the semiannual oscillation in the Baltic, Red, Black, Caspian, east Mediterranean Seas and Persian Gulf. During the El Niño 1997/98 an increase of the amplitude of the semiannual oscillation can be noticed in the east equatorial Pacific Ocean together with simultaneous decrease of its amplitude in the Indian Ocean. The amplitude of the semiannual oscillation diminish in all the European seas, Red Sea and Persian Gulf in 1999 during La Niña.

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References


Kosek W. 1999a, Analysis of short period oscillations of sea level anomalies computed from TOPEX/Poseidon altimetry, Report No 40, Space Research Centre, Polish Academy of Sciences.


PSMSL 2001, Permanent Service for Mean Sea Level Training Information, Reports and Manuals etc. http://www.pol.ac.uk/psmsl/training/training.html


