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RESEARCH ARTICLE

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Key Points:

- Expected global sea level rise can be monitored with few years of data
- At regional scale, the required period is much longer
- Decadal variability determines the regions with larger detection time

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Detection time for global and regional sea level trends and accelerations

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Abstract Many studies analyze trends on sea level data with the underlying purpose of finding indications of a long-term change that could be interpreted as the signature of anthropogenic climate change. The identification of a long-term trend is a signal-to-noise problem where the natural variability (the “noise”) can mask the long-term trend (the “signal”). The signal-to-noise ratio depends on the magnitude of the long-term trend, on the magnitude of the natural variability, and on the length of the record, as the climate noise is larger when averaged over short time scales and becomes smaller over longer averaging periods. In this paper, we evaluate the time required to detect centennial sea level linear trends and accelerations at global and regional scales. Using model results and tide gauge observations, we find that the averaged detection time for a centennial linear trend is 87.9, 76.0, 59.3, 40.3, and 25.2 years for trends of 0.5, 1.0, 2.0, 5.0, and 10.0 mm/yr, respectively. However, in regions with large decadal variations like the Gulf Stream or the Circumpolar current, these values can increase up to a 50%. The spatial pattern of the detection time for sea level accelerations is almost identical. The main difference is that the length of the records has to be about 40–60 years longer to detect an acceleration than to detect a linear trend leading to an equivalent change after 100 years. Finally, we have used a new sea level reconstruction, which provides a more accurate representation of interannual variability for the last century in order to estimate the detection time for global mean sea level trends and accelerations. Our results suggest that the signature of natural variability in a 30 year global mean sea level record would be less than 1 mm/yr. Therefore, at least 2.2 mm/yr of the recent sea level trend estimated by altimetry cannot be attributed to natural multidecadal variability.

1. Introduction

There are many studies analyzing linear trends on sea level data with the underlying purpose of finding indications of anthropogenic climate change. The basic assumption implicitly adopted is that the anthropogenic effects on the climate would be reflected as a long-term (i.e., centennial or longer) change. Despite the possible presence of nonlinear trends [e.g., Jevrejeva *et al.*, 2008; Woodworth, 2006; Merrifield *et al.*, 2009; Church *et al.*, 2013], a linear trend is often used to quantify sea level change at both global and regional scales. Linear trend estimates are obtained from coastal tide gauges or altimetry data over a wide range of time spans, from few years to centennial scales [e.g., Nerem, 1995; Cabanes *et al.*, 2001; Church and White, 2011; Cazenave and Llovel, 2010; Spada and Galassi, 2012]. However, any eventual long-term trend (e.g., from anthropogenic origin but not necessarily) is embedded in the background “noise” of natural climate variability [e.g., Hasselmann, 1979; Santer *et al.*, 2000; Rybski *et al.*, 2006], which can distort the detection of forced trends over rather long periods of time [Wunsch, 1999; Hoerling *et al.*, 2010; Franzke, 2012]. This problem has been particularly acknowledged in the detection of centennial sea level trends, due to the significant interannual and multidecadal variability of sea level records [Barbosa *et al.*, 2008; Landerer *et al.*, 2008; Hamlington *et al.*, 2011; Zhang and Church, 2012; Calafat and Chambers, 2013].

In practice, the identification of a forced linear or quadratic trend in an observational record is a signal-to-noise (S/N) problem where the natural variability (the “noise”) can mask the externally forced trend (the “signal”). This S/N ratio depends on the magnitude of the forced trend or acceleration, on the magnitude of the natural variability, and on the length of the record, as the climate noise is larger when averaged over short time scales and becomes smaller over longer averaging periods [Santer *et al.*, 2011]. In this context, the arising question is how long should a sea level record be to allow the identification of a linear or quadratic long-term trend (i.e., centennial) with a reasonably degree of confidence? In other words, what is the detection time (DT) of sea level trend and accelerations? We can also formulate the question the other way: if we

have a sea level record of a certain length, what is the minimum trend or acceleration magnitude that can be reliably identified from the record? This information is important to reinforce the confidence on the conclusions related to anthropogenically induced climate change and to provide policy makers with appropriate information [GCOS, *Climate Monitoring Principles*, 2013]. It is important to clarify here that the detection of a long-term (i.e., centennial) trend is not enough to conclude that it is a forced trend. Whether such trend is of anthropogenic origin or not would require a specific attribution study.

Douglas [1991] addressed the detection issue in a qualitative way by comparing the trend values obtained from tide gauges of varying length. Hughes and Williams [2010] used a more quantitative approach based on the modeling of sea level natural variability as a high-order autoregressive process. However, their results were based on 12 years of altimetry data which, as we will show later, is a strong limitation. Concerning sea level accelerations, Woodworth [1990] discussed the detection time of sea level acceleration at the Newlyn tide gauge, using a simple characterization of the natural variability. In a recent work, Haigh *et al.* [2014] analyzed the detection time of acceleration in sea level rise at several tide gauge locations and for the GMSL. They defined the DT as the time when the computed acceleration is significantly different from zero independently of its value. They computed the statistical significance considering that the sea level natural variability can be modeled as a first-order autoregressive process. However, Hughes and Williams [2010] have suggested that to model sea level variability as an autoregressive process it should be of fifth order. Thus, the results of Haigh *et al.* [2014] may be biased low.

In this paper, we attempt to determine the detection time of sea level linear trends and accelerations at global and regional scales using tide gauge, altimetry and numerical model data to characterize the natural variability. The DT time is here defined as the period required to identify a centennial trend with a reasonably degree of confidence (here set to a 20% of the actual value). We do this for different magnitudes of the linear trends and accelerations representative of observed and projected sea level changes for the 21st century.

2. Data

Monthly time series of sea level observations from tide gauges were obtained from the Revised Local Reference data archive of the Permanent Service for Mean Sea Level [Woodworth and Player, 2003]. We selected 15 long (>80 years) tide gauge records with almost continuous data (less than 10% of gaps). The inverse barometer correction was applied using atmospheric sea level pressure from the Twentieth Century Reanalysis project [Compo *et al.*, 2011]. No correction for land movements has been applied as far as linear trends will be removed from each record before using it (see below).

The global mean sea level (GMSL) monthly values computed for the period 1900–2000 by Jevrejeva *et al.* [2008], Church and White [2011], and Calafat *et al.* [2014] have been used to characterize the natural variability of GMSL. The first reconstruction is obtained with the virtual station method, while the others are obtained with an EOF-based method (Reduced-Order Interpolation, ROI). Moreover, Calafat *et al.* [2014] provide two estimates of GMSL. The first one is obtained using the same method as Church and White [2011] but using a different set of tide gauges, while in the second one GMSL is reconstructed without including the EOF-0 (a spatially constant mode). Those authors have shown that the ability of the ROI method to reproduce the interannual variability is significantly degraded when the EOF-0 is included in the reconstruction, a feature that is very relevant to the problem studied here. Additionally, we also consider the monthly global maps obtained from a regional reconstruction as provided by Church *et al.* [2004].

Altimetry maps of sea level anomalies for the period (1993–2012) were collected from the Archiving, Validation and Interpretation of Satellite Oceanographic data service (AVISO; <http://www.aviso.oceanobs.com>). All the standard corrections (tides, wet/dry troposphere, ionosphere, and atmospheric forcing) were applied. The weekly $1/4^\circ \times 1/4^\circ$ maps were monthly averaged for consistency with the other data sets.

Monthly sea level fields from several global ocean models have been used to cover a longer period than the one provided by altimetry. The simulations have different characteristics in terms of numerics, forcings, spatial resolution, and time coverage. We have used ORAS-4 [Balmaseda *et al.*, 2013] (1° resolution, 1958–2012 period), SODAv2 [Carton and Giese, 2008] (0.5° resolution, 1871–2008 period), ORCA-G70 [Penduff *et al.*, 2010] ($1/4^\circ$ resolution, 1958–2004 period), and GECCO [Köhl and Stammer, 2008] (1° resolution, 1952–2001 period). SODAv2, ORAS-4, and GECCO assimilate temperature and salinity profiles and satellite sea

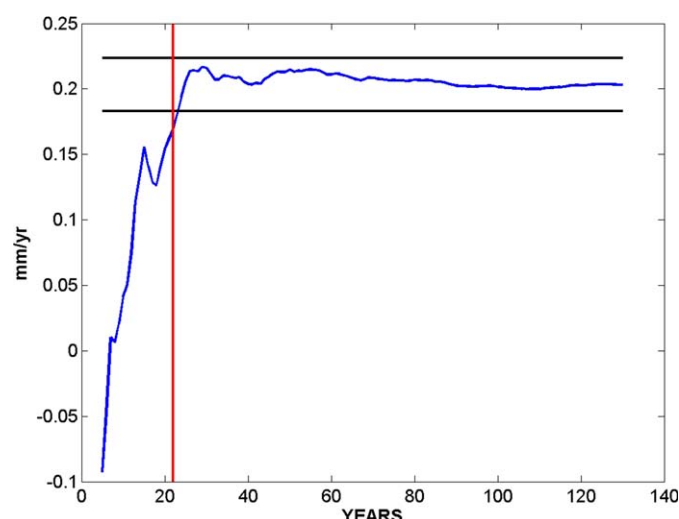


Figure 1. Example of the linear trend obtained from a surrogate time series of Global Mean Sea Level (GMSL) using different record lengths. The black lines indicate the $\pm 20\%$ of the actual trend and the red line the time when this accuracy is reached.

particular, we use a modified version of the phase scrambling method of *Theiler et al.* [1992]. In this method, the power spectrum of a time series is computed and the surrogate time series is obtained recombining the different components of the spectrum but with random phases. For all time series used here and for the fields of the oceanic models, we ensure that the phases corresponding to the same peak of energy have the same random perturbation, so that coherent signals are not distorted. Using this phase scrambling method, the surrogate time series has the same autocorrelation function as the original time series. In our case, all the synthetic time series (that is, those generated from different GMSL estimates, from altimetry data, from tide gauge records, and from model data) are 100 years long regardless of the period spanned by the original series.

The second step is to add a known linear trend (*target trend*) to the synthetic time series, so that the actual underlying trend of the series is perfectly known. Then, we estimate the linear trend using subperiods of varying length, from 10 to 100 years (see the blue curve in Figure 1), and we define the DT as the record length required to get an estimate of the linear trend which only differs from the actual value in a 20% (the red line in Figure 1). This procedure is repeated for N different synthetic time series (surrogates) to produce a distribution of $\{DT^i\}$ values. In this study, N has been set to 500; using larger values does not significantly change the results. For illustrative purposes, different values of the target trend are used: 0.5, 1, 2, 5, and 10 mm/yr. The 0.5 and 1 mm/yr are representative of the lower bound of observed regional trends. Two millimeter per year is close to the estimated GMSL trend for the twentieth century [*Jevrejeva et al.*, 2008; *Church and White*, 2011]. Five millimeter per year would correspond to a moderate sea level rise scenario for the 21st century as obtained from process-based models [*Church et al.*, 2013], while 10 mm/yr would represent the same quantity but obtained from semiempirical models [*Rahmstorf*, 2007].

An analogous procedure has been applied to determine the DT of sea level accelerations. In this case, surrogate series are created adding a known acceleration (*target acceleration*) to the synthetic time series. No linear term is added, so the long-term change is only due to the acceleration. The values used for the target acceleration are: 0.025, 0.05, 0.075, and 0.1 mm/yr², which correspond to a change of 25, 50, 75, and 100 cm in 100 years, respectively. The 0.025 mm/yr² is representative of the observed accelerations at tide gauges during the twentieth century [e.g., *Calafat and Chambers*, 2013; *Haigh et al.*, 2014]. The 0.05 and 0.075 mm/yr² are representative of moderate and high sea level rise scenarios as obtained from process-based models [*Church et al.*, 2013], while 0.1 mm/yr² would represent a moderate sea level rise obtained from semiempirical models [*Rahmstorf*, 2007].

4. Results

4.1. Detection Time of Global Mean Sea Level Trends

The distribution of DT obtained for each target trend using GMSL data from different sources is represented as a box plot in Figure 2a. The spread of DT values is noticeable in all the cases, with an interquartile range

surface temperature. ORAS-4 and GECCO also assimilate altimetry data. ORCA-G70 is a forced run without data assimilation. All the models are forced by atmospheric reanalysis fields and none of them include the inverse barometer effect on sea level.

3. Methods

For each time series considered, the DT corresponding to a given target trend is estimated using the following procedure. First, the time series is detrended and the annual cycle is removed (although the latter has no significant impact on the result). Then, a nonparametric method is used to create surrogate data. In par-

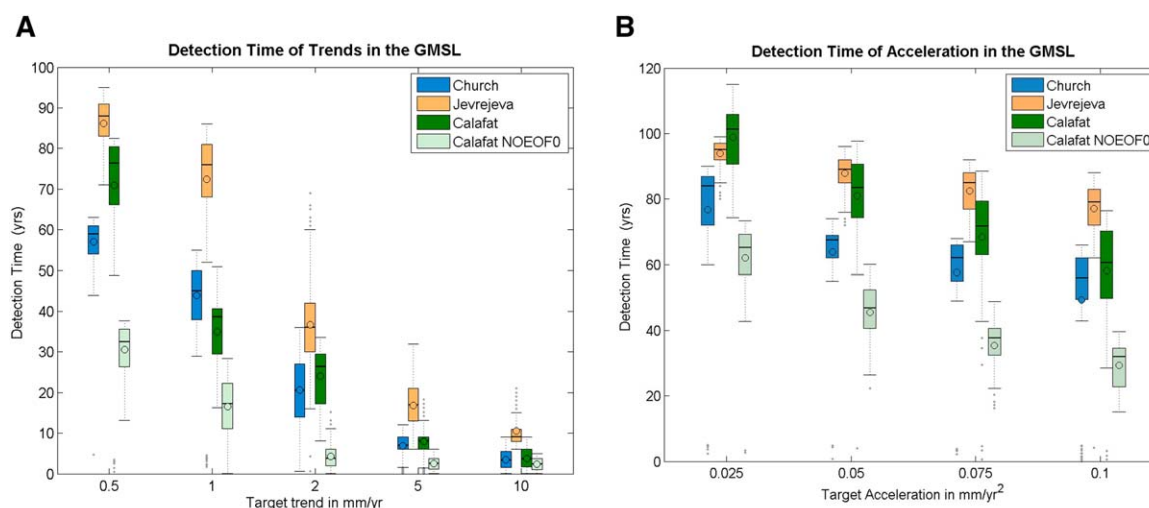


Figure 2. (a) Box plot of the Detection Time (DT) corresponding to different target trends obtained using different GMSL reconstructions. (b) The same but for the DT corresponding to different target accelerations. The box bounds in the box plots indicate the 25 and 75 percentile of the DT distribution while the whiskers indicate the 5 and 95 percentiles.

of about 10 years. In the following, we will refer to the median value of the DT distributions. From Figure 2a, it can be seen that the DT decreases as the target trend increases, as expected. Using the GMSL from *Church and White* [2011], the DT for a trend of 0.5 mm/yr is 59 years and decreases to 45, 20, 10, and 3 years if the target trend is 1, 2, 5, and 10 mm/yr, respectively. The other estimates of GMSL lead to a similar behavior, although the DT values for each target trend can be very different. In particular, the GMSL from *Jevrejeva et al.* [2008] leads to the largest values for the DT, while the GMSL obtained without the EOF-0 [*Calafat et al.*, 2014] leads to the lowest values. This is in good agreement with the fact that the interannual variability, quantified in terms of the standard deviation (std) of the series, is much larger in *Jevrejeva et al.* [2008] (std = 1.4 cm) than in *Church and White* [2011] (std = 0.8 cm) and *Calafat et al.* [2014] (std = 0.9 cm), and with the fact that the variability without the EOF-0 is much reduced (std = 0.2 cm). It must be noted that the discrepancies among data sets do not reflect uncertainties in the estimate of the DT: they are the consequence of the differences in the strength of the interannual and interdecadal variability in each GMSL series. In other words, there is not a “correct” value for the DT; it depends on the representation of the natural variability in each time series.

Additionally, a note of caution needs to be sounded. The *Calafat et al.* [2014] reconstruction without the EOF0 is the one showing the more realistic interannual variability and the lowest DT. Thus, one could be tempted to think that this reconstruction would be the best choice to compute GMSL centennial trends. However, *Calafat et al.* [2014] have acknowledged that a reconstruction without the EOF0 is not able to capture centennial trends or accelerations.

4.2. Detection Time of Global Mean Sea Level Accelerations

The distribution of DT obtained for each target acceleration and using the different GMSL reconstructions is represented in Figure 2b. There is again a spread of DT values in all the cases, with an interquartile range of about 15–20 years. In general, the DT of the accelerations is larger than the DT of a linear trend that would produce the same change in 100 years. For instance, the DT of a 0.05 mm/yr² acceleration is between 45 and 85 years, while the DT for a 5 mm/yr trend is between 5 and 15 years. This is because, during the first years of the record, a quadratic term only induces small changes, which are difficult to detect in the presence of natural variability. Toward the end of the record, the changes induced by the quadratic term are much stronger and it is then when the acceleration can be detected.

The different GMSL reconstructions lead again to different values for the DT of the accelerations. The reconstruction of *Jevrejeva et al.* [2008] is the one leading to the larger values with a DT of 95, 90, 85, and 80 years for a target acceleration of 0.025, 0.05, 0.075, and 0.1 mm/yr². Again, the GMSL obtained without the EOF-0 [*Calafat et al.*, 2014] leads to the lowest values (65, 45, 38, 30 years, respectively).

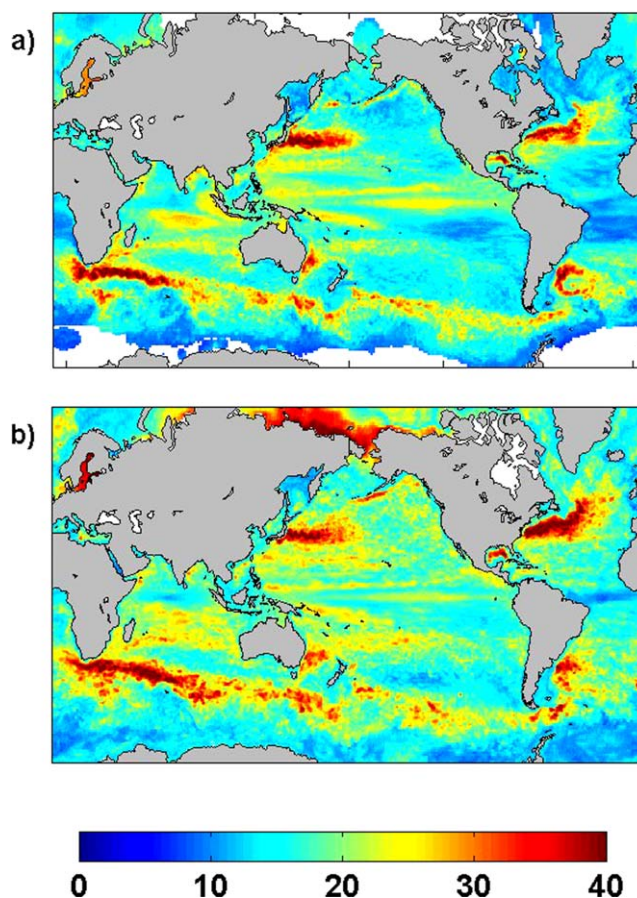


Figure 3. Detection time maps for a target trend of 2 mm/yr obtained using (a) the variability inferred from the altimetry data set (1993–2012); (b) the variability inferred from a subset of the ORAS4 model output spanning the altimetric period.

4.3. Detection Time of Regional Sea Level Trends

The DT of regional sea level trends have been first estimated using surrogate time series of altimetry data. The same methodology as used for the global sea level time series has been used for each point of the altimetry field. That is, the variability of the 20 year records of altimetry data at each grid point has been used to generate 100 year time series at the same grid point. The DT map for a target trend of 2 mm/yr is shown in Figure 3a. The DT distribution is not homogeneous, being larger in regions of strong dynamics or affected by large interannual or interdecadal variability. While values range between 5 and 15 years over most of the open ocean, they increase up to 40–50 years in the regions of the Gulf Stream, the Kuroshio, and the Agulhas current. Relatively high values (~30 years) are also obtained in the Southern ocean associated with the circumpolar current and also in the tropical Pacific, where El Niño/La Niña signature on sea level is especially strong. These results are very similar

to those shown by *Hughes and Williams* [2010, Figure 7b] using a fifth-order autoregressive model to characterize the natural variability. The global averaged value of the DT for this target trend is 17.6 years with a standard deviation of 7.3 years (Table 1). The global average of the DT ranges from 36.4 to 7.3 years, for target trends between 0.5 and 10 mm/yr.

For comparison, we include also the DT map obtained with the ORAS4 data but using only the variability of the altimetric period (1993–2012; Figure 3b) to construct the surrogates. The spatial patterns are in very good agreement with the patterns obtained from altimetry. The values of the DT in the regions of high sea level variability are also well captured and only in the places where altimetry showed the lowest values ORAS4 is slightly biased high (5–10 years). Additional computations have been done using other 20 year periods from the ORAS4 data set, but the DT maps obtained were very similar to those shown in Figure 3b.

We now compute the DT using the whole period of the ORAS4 simulation (1958–2012) to generate the 100 year time series. The key feature to be noted is that the use of a longer period to characterize the sea level

Table 1. Global Average of the Detection Time (in Years) \pm the Standard Deviation Corresponding to Several Target Trends and Obtained Using Different Sea Level Data Sets

Target Trend (mm/yr)	AVISO	ORAS4	SODAv2	ORCAG70	GECCO	CHURCH 04 Reconst.	AVISO + 60 yr Oscillation
0.5	36.4 \pm 14.9	87.9 \pm 38.8	85.8 \pm 38.0	78.0 \pm 40.6	90.4 \pm 26.6	60.8 \pm 23.8	95.0 \pm 24.5
1.0	25.3 \pm 10.5	76.0 \pm 35.4	71.4 \pm 34.7	65.8 \pm 36.5	79.6 \pm 25.8	45.8 \pm 20.1	88.9 \pm 23.0
2.0	17.6 \pm 7.3	59.3 \pm 28.5	53.6 \pm 26.5	50.0 \pm 28.2	63.2 \pm 21.7	31.3 \pm 17.9	73.9 \pm 19.7
5.0	10.8 \pm 4.5	40.4 \pm 21.6	36.3 \pm 20.1	32.8 \pm 20.5	43.2 \pm 16.0	14.0 \pm 12.6	51.3 \pm 13.4
10.0	7.3 \pm 3.6	25.2 \pm 17.3	22.3 \pm 16.0	20.7 \pm 16.3	29.5 \pm 15.2	6.2 \pm 6.6	36.4 \pm 9.7

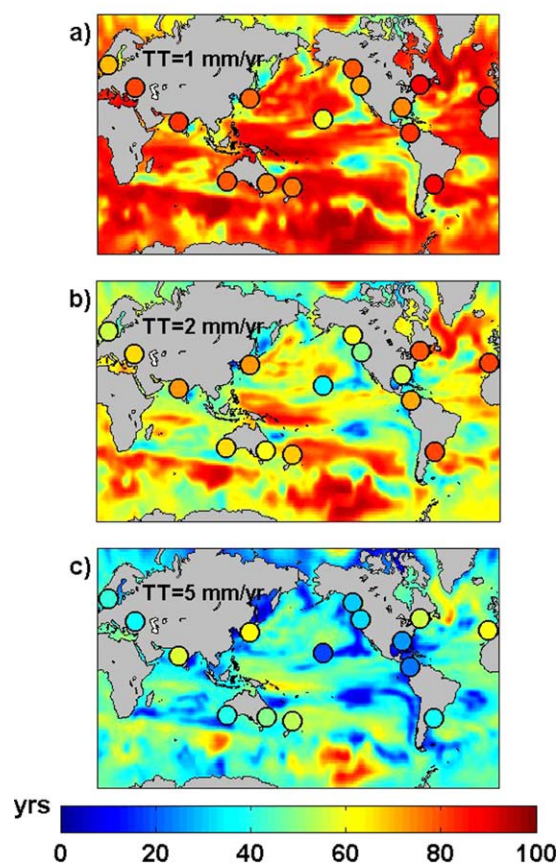


Figure 4. Detection time (DT) maps obtained using the variability inferred from the whole ORAS4 model output (1958–2012). The colored dots show the DT obtained using the variability inferred from tide gauges. They have been computed for different target trends: (a) 1 mm/yr, (b) 2 mm/yr, (c) 5 mm/yr.

have detected a 60 year signal with amplitudes ranging from 0.5 to 2.5 cm in several coastal tide gauge records. The existence of such signal in the open ocean is still unknown mainly due to the lack of suitable data to confirm or reject that hypothesis. Nevertheless, for illustrative purposes, we have artificially added a 60 year harmonic signal of 2 cm amplitude to each grid point of the altimetric data set. The result is a significant increase of the DT with respect to the values obtained using only altimetry data (Table 1). Namely, the existence of such a relatively small signal increases the DT by a factor of 3–5, so that the global average of the DT ranges between 95.0 and 36.4 years for the different target trends.

The spatial patterns obtained from the whole ORAS4 data set have also changed with respect to those obtained from altimetry (compare Figures 4b and 3a). This is again due to the fact that the whole ORAS4 data set includes information on the decadal variability and, as already pointed out by *Zhang and Church* [2012], the interannual and decadal sea level patterns are not necessarily coincident. Some of the features shown in Figure 4 are consistent with previous studies on decadal sea level variability. The high values of the DT in the eastern North Atlantic may be related to decadal variability associated with large-scale atmospheric changes [*Sturges and Douglas*, 2011] and, in particular, with longshore winds [*Calafat et al.*, 2012]. *Häkkinen* [2001] also reported decadal sea level variability along the Atlantic western boundary and linked it to changes in the basin-scale thermal forcing and to overturning changes. In the Indian Ocean, *Timmermann et al.* [2010] and *Nidheesh et al.* [2013] have shown the role of wind on sea level decadal variability. Decadal zonal wind stress variations induce steric sea level fluctuations in the eastern equatorial Indian Ocean and the Bay of Bengal. Wind stress curl in the southern Indian Ocean drives decadal variability in the south-western Indian Ocean [*Nidheesh et al.*, 2013]. Variations of the easterly winds in the central and western Pacific Ocean

variability leads to a large increase in the DT: the global average of the DT is now 87.9, 76.0, 59.3, 40.3, and 25.2 years for a target trend of 0.5, 1.0, 2.0, 5.0, and 10.0 mm/yr, respectively (Table 1). The reason is that the 20 years of the altimetric period is too short to properly characterize the decadal variability and hence to distinguish which part of the signal is due to decadal variability and which part is due to a long-term trend. Thus, when time series are detrended prior to the spectral analysis, part of the decadal variability is removed. This means that the surrogate data generated from altimetry do not contain the whole decadal signal, which in turn makes easier the detection of linear trends. Conversely, when the whole period of ORAS4 data is used, the detrending does not remove the interdecadal signal, and the surrogate data created from ORAS4 include decadal variability. This means that, in practice, the decadal variability can contaminate the trend computation if the record is not long enough, even if it has a small amplitude compared to the interannual variability. Conversely, seasonal and interannual variability have a small impact on the trend estimates when the record length is longer than 10–20 years, even if they have large amplitudes compared to longer-term variability.

As an example of the role of the decadal variability on the DT, we have performed an additional experiment. *Chambers et al.* [2012]

Table 2. Comparison of the Detection Time Obtained Using Tide Gauge Observations With the Values Obtained Using Different Models at the Tide Gauge Locations^a

Target Trend (mm/yr)	ORAS4	ORCAG70	SODAv2	GECCO	CHURCH 04 Reconst.
BIAS < Model – Observations >					
0.5	–1.5	–5.1	–5.2	–0.3	–19.9
1.0	–6.4	–10.4	–12.1	–5.3	–26.7
2.0	–10.0	–18.0	–15.6	–9.7	–28.0
5.0	–5.6	–9.8	–9.3	–4.5	–24.5
10.0	–6.2	–8.6	–8.8	–4.3	–19.7
RMSE					
0.5	9.4	14.8	15.4	11.0	23.4
1.0	14.2	18.5	21.7	18.0	28.8
2.0	16.2	23.9	23.6	20.0	31.3
5.0	11.9	21.0	24.3	17.6	28.6
10.0	15.8	16.6	20.2	14.3	22.5

^a(Top) Averaged Bias in years. (Bottom) RMS error in years.

are also relevant. The strong decadal wind stress curl fluctuations on the northern and southern flanks of the western Pacific force Rossby waves, which are responsible for decadal sea level variations in the northwest and southwest Pacific regions [Sasaki *et al.*, 2008; Nidheesh *et al.*, 2013].

The details of the DT patterns show some dependence on the numerical model. Thus, in order to validate our results, we have also computed the DT using tide gauge records (see colored dots in Figure 4) and compared them with the values obtained from ORAS4 at the closest grid point to each tide gauge. There is an

overall good agreement between the two sets of results, both in the magnitude and in the regional patterns (the two of them reproduce the differences between the eastern and western coasts of Australia or the relative minimum in the subtropical Pacific, for instance). However, the model DTs show a negative bias with respect to tide gauge DT (between 1.5 and 10 years depending on the target trend, see Table 2). The RMS error between both sets of results ranges between 9.4 and 16.2 years depending on the target trend.

Additionally, and in order to check the robustness of model-derived results, the DT has also been derived from other models. The global average of the DT is in good agreement among models, with GECCO providing the largest values and ORCAG70 the smallest. In all cases, the typical difference between models is around 5 years (Table 1). The spatial patterns are also in overall good agreement among models (not shown). Compared with the values obtained from tide gauges, GECCO shows a small bias and RMS errors close to those obtained from ORAS4 while ORCAG70 and SODAv2 show similar performance with larger biases and RMS errors.

Finally, we have computed the DT from the sea level maps reconstructed by Church *et al.* [2004]. They are a factor 1.5–5 smaller than those

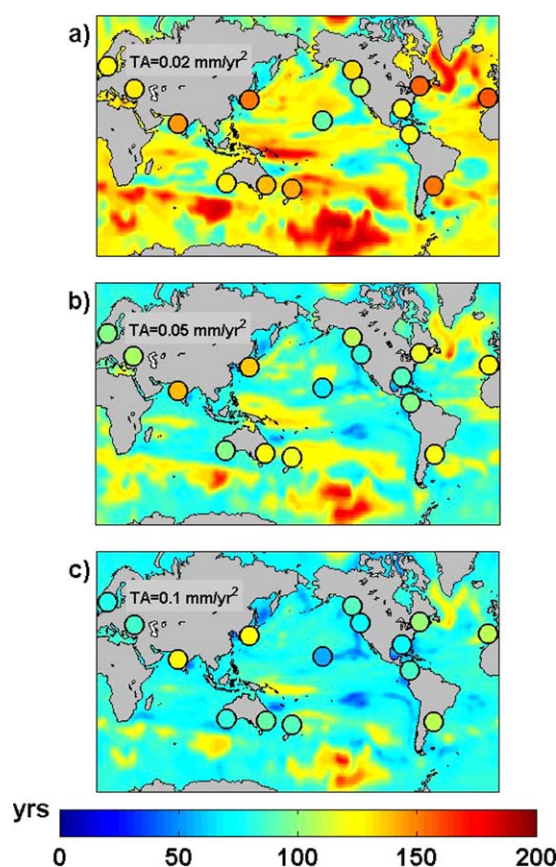


Figure 5. Like Figure 4 but for the detection time (DT) of different accelerations: (a) 0.02 mm/yr², (b) 0.05 mm/yr², (c) 0.1 mm/yr². Note the different colorbar with respect to Figure 4.

Table 3. Global Average of the Detection Time (in Years) \pm the Standard Deviation Corresponding to Several Target Accelerations and Obtained Using Different Sea Level Data Sets

Target Acceleration (mm/yr ²)	ORAS4	SODAv2	ORCAG70	GECCO	CHURCH 04 Reconst.
0.025	119.3 \pm 55.6	110.4 \pm 52.6	100.6 \pm 41.2	126.5 \pm 40.3	78.4 \pm 26.9
0.050	92.8 \pm 43.8	86.1 \pm 40.8	80.9 \pm 32.9	99.5 \pm 33.2	63.9 \pm 22.8
0.075	81.0 \pm 36.3	75.5 \pm 34.1	71.9 \pm 28.4	85.9 \pm 26.7	54.9 \pm 23.3
0.100	75.3 \pm 33.6	70.2 \pm 31.5	66.4 \pm 26.0	78.1 \pm 22.8	47.6 \pm 23.1

obtained with the models (Table 1). The comparison with tide gauges shows the largest biases and RMS errors (Table 2). This suggests that the regional decadal variability of the *Church et al.* [2004] reconstruction is underestimated. At the time of this reconstruction was generated, only 8 years of data were available from altimetry to compute basis functions, so the EOF base used to reconstruct the sea level field may be not rich enough. This could explain the problems of the *Church et al.* [2004] reconstruction to capture inter-annual and especially multidecadal variability at regional scale.

4.4. Detection Time of Regional Sea Level Accelerations

The regional distribution of the DT for different sea level accelerations and computed from the ORAS4 data set is presented in Figure 5. The spatial patterns are almost identical to those obtained for the DT of linear trends (compare with Figure 4). However, the magnitude of the DT is larger. The averaged value of the DT is 119.3 ± 55.6 , 92.8 ± 43.8 , 81.0 ± 36.3 , and 75.3 ± 33.6 years for target accelerations of 0.025, 0.050, 0.075, and 0.100 mm/yr², respectively. As noted above, for a given sea level change, a longer record is required if the change follows a quadratic function than if it follows a linear trend. In particular and for the range of changes analyzed here, the length of the record must be ~ 40 – 60 years longer to detect an acceleration than to detect a linear trend producing an equivalent sea level change after 100 years.

The DT obtained from tide gauge data again compares well with the results obtained from the ORAS4 data set. Although there are some discrepancies, the spatial heterogeneity of the DT shown by the tide gauges is properly reproduced by ORAS4 (Figure 5). Furthermore, the global average of the DT of sea level accelerations is also in good agreement among the different models analyzed here with GECCO providing the largest values and ORCAG70 the smallest. The typical difference between models is around 10 years (Table 3).

5. Discussion and Conclusions

The time required to get robust estimates of centennial linear trends and accelerations of regional and global mean sea level has been estimated using observational and numerical data sets. The DT is larger when/where the actual trend or acceleration is smaller or when/where the underlying interdecadal variability is larger. A centennial linear trend in the global mean sea level of 5 mm/yr, for instance, which is representative of the projected sea level rise during the 21st century under a moderate climate change scenario, could be identified using 5–15 years of data. However, identifying the same trend from regional-scale observations would require, on average, a minimum period of 40 years. This value would increase up to 60–80 years in regions with strong decadal sea level variability such as the circumpolar current. The spatial pattern of the DT of sea level accelerations is almost identical to the pattern corresponding to the DT of sea level trends. The main difference, however, is that the time required to detect accelerations is much larger. The length of the records has to be about 40–60 years longer to detect an acceleration than to detect a linear trend resulting in an equivalent change after 100 years.

All the models analyzed in this paper provide similar values and are also in a reasonably good agreement with the values derived from tide gauge observations. This good agreement means that the spectral content of the different data sets is similar, specially at low frequencies (i.e., multidecadal periods). Also, an important point is that the determination of the DT cannot be undertaken using only a few decades of data (e.g., the period spanned by altimetry). At least several decades are needed to properly characterize the decadal variability and hence to ensure that the generated surrogate time series reflect the actual variability. This explains why the experiments carried out using altimetry data lead to a large underestimation of the DT. Another point is that the data used to characterize the variability must provide a good

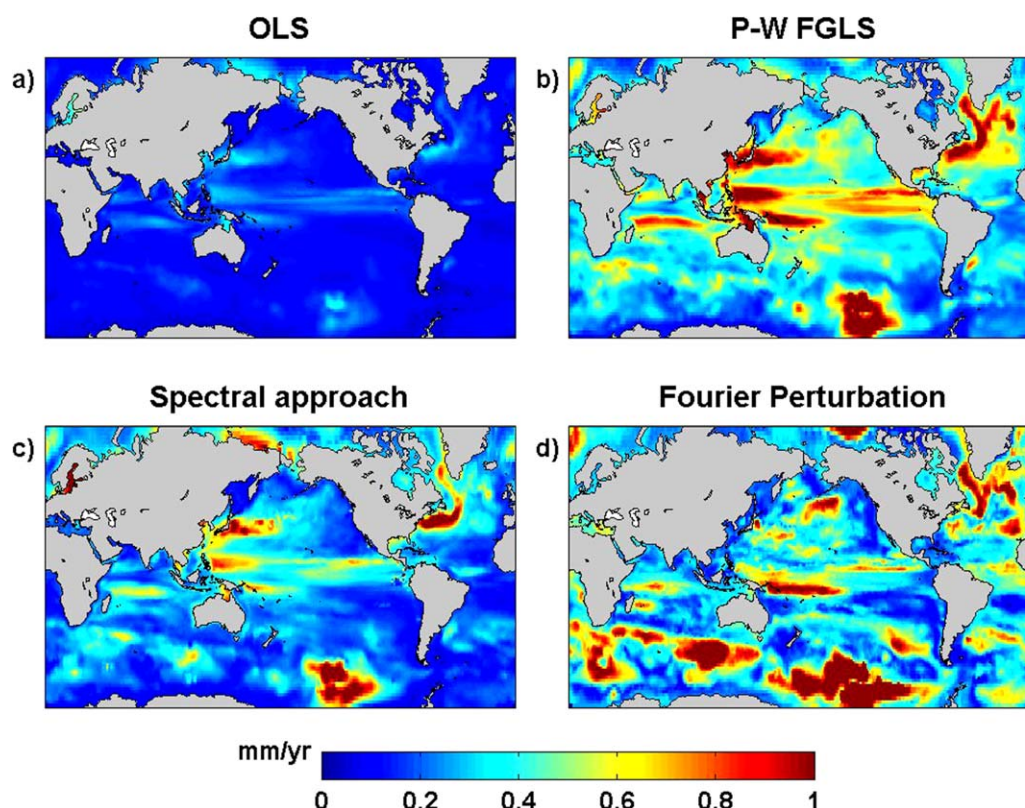


Figure 6. Standard error of the linear regression for a 50 years record of the ORAS4 data set as estimated with different methods. (a) Ordinary Least Squares, (b) feasible generalized least squares using Prais-Winsten estimation, (c) spectral approach, and (d) Fourier perturbation. See the text for details.

representation of the actual variability, regardless of the spanned period. The regional sea level reconstruction of *Church et al.* [2004], for instance, leads to small DT values probably because the reconstruction underestimates the interannual and decadal variability at regional scale.

The main consequence of this study is that special care must be paid when interpreting linear trends and/or accelerations obtained from short records (e.g., from altimetry) and/or from local measurements (e.g., from tide gauges). Although this is somehow an obvious conclusion, there are many examples in the literature where the trends or accelerations are obtained from short records (e.g., altimetry) and implicitly interpreted as potential signs of climate change. Thus, the DT values shown here can be used as a first indicator to judge if a certain measured trend or acceleration is representative of a long-term (i.e., centennial) trend or if, conversely, it cannot be distinguished from the natural variability (i.e., interannual to multidecadal). As a second step, it would be recommended that the statistical significance of the trends was computed considering the existence of long-term memory in the records. This is not straightforward but there are different possibilities proposed in the literature [see, for instance, *Barbosa et al.*, 2008, and references therein]. As an example, we have compared the standard error of the linear trend obtained using four different methods. First, we use the standard error provided by the ordinary least squares, which assumes that residuals are decorrelated (i.e., no-memory) and which is often the approach followed in the literature. Then, we use the error provided by a feasible generalized least squares using Prais-Winsten estimation. The third method tested here is the one proposed by *Smith* [1993] where the long-range dependence is estimated by a spectral-based method. Finally, we apply the same approach followed here to compute the DT to generate the standard error of the trend computation. The four methods have been applied to the ORAS4 sea level data set (Figure 6). The globally averaged value of the standard error using ordinary least squares is 0.14 mm/yr while it is 0.44, 0.30, and 0.39 mm/yr when using the generalized least squares, the spectral-based method, or the frequency perturbation methods, respectively. Therefore, it is clear that ordinary least squares clearly underestimate the standard error. Concerning the methods which consider long-term dependence, there are some differences among them both in terms of magnitude and spatial pattern but

in general they are in good agreement. In consequence, if the goal is to identify long-term trends it would not be recommended to estimate the trend significance using ordinary least squares.

Once it has been demonstrated that (for a long enough record) the most influential factor determining the DT of centennial trends is the amplitude of the low frequency (i.e., from multiannual to multidecadal) variability, it seems clear that removing this variability from the sea level records would help to reduce the DT and hence to increase the reliability of the identification of long-term trends. In principle, this could be done using information about the dominant processes driving the decadal variability [Zhang and Church, 2012; Calafat and Chambers, 2013; Cazenave *et al.*, 2014] or using mathematical tools to improve the signal to noise ratio [Hamlington *et al.*, 2011]. However, both options require a good knowledge on the characteristics of sea level decadal variability, which is rather poor in the open ocean and in many coastal regions. Moreover, there are noticeable discrepancies among results inferred from numerical models in some regions. This points to maintaining in time the altimetric observing systems as a primary tool to characterize the decadal signals and hence to increase the reliability in the detection of sea level long-term trends and accelerations at regional scale.

A final remark is referred to the DT for the global mean sea level and the differences in the results depending on the sea level reconstruction chosen to characterize the GMSL. Calafat *et al.* [2014] have shown that a sea level reconstruction without the EOF-0 leads to a much better interannual variability. Moreover, they have shown that the amplitude of the variability is also close to the amplitude measured by altimetry while the reconstruction which does include the EOF-0 shows a variability of about 4–5 times larger than altimetry. This has important consequences for the attribution of the GMSL recent trends measured with altimetry. For the period 1993–2010, the observed trend is 3.2 mm/yr [e.g., Nerem *et al.*, 2010]. If the natural variability is characterized from a reconstruction including the EOF-0 [e.g., Church and White, 2011], then a 30 year long record is not enough to ensure with a 95% confidence that a 3 mm/yr is not due to natural variability (see Figure 2a). However, if the natural variability is characterized from a reconstruction which does not include the EOF-0 [e.g., Calafat *et al.*, 2014] then the DT for a 3 mm/yr trend reduces to less than 10 years. In consequence, as the variability produced by such reconstruction is much closer to the real variability, it seems safe to state that the GMSL trend observed from altimetry is, in a large fraction, not induced by natural variability. In other words, the signature of natural variability in a 30 year record (i.e., that of altimetry) would be less than 1 mm/yr. Therefore, at least 2.2 mm/yr of the recent sea level trend estimated from altimetry cannot be attributed to natural interannual or multidecadal variability.

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Erratum

In the originally published version of this article, Figure 2b was incorrectly typeset as Figure 2a, and Figure 2b did not show the correct image. The error has since been corrected and this version may be considered the authoritative version of record.