Small Publications in Historical Geophysics

No. 7

Determination of Global Sea Level Rise and its Change with Time

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Summer Institute for Historical Geophysics
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1. Introduction

It is well known today that sea level is globally rising due to a comparatively mild climate with melting glaciers and other effects. This was first detected by Thorarinsson (1940) and Gutenberg (1941). However, what has turned out to be difficult to determine is the rate at which the sea level rises, and its possible changes with time. Even more difficult to find out is to what extent the rise of the sea level is caused by a natural climatic variation and to what extent it is caused by a man-made greenhouse effect. In order to understand the origin of the sea level rise and, thereby, perhaps also its future development, it is of great importance to determine both past and present rates of the sea level rise.

There are two basic problems connected with the determination of the global (eustatic) sea level rise. First, the sea level stations (tide gauges, mareographs) are, or at least should be, fixed in the crust. Hence also possible vertical motions of the crust are recorded in the sea level data. Second, the sea level stations cover different time spans, most of them not very long ones. Therefore the data also might contain sea level changes of other, more short-term, origins than the global one. In addition, the geographical distribution of the sea level stations over the globe is very uneven. Thus they hardly form any representative set of stations.

To try to overcome these complications, various methods for determining the general climatic sea level rise have been developed. They can be loosely described as follows:

- Global averaging or series expansion of sea level trends.
- Comparing postglacial rebound models with sea level trends.
- Comparing vertical crustal motion recordings with sea level trends.
- Investigating changes with time of extremely long sea level trends.

The sea level trend itself at a certain station is calculated by ordinary linear regression of the time series of observed annual mean sea levels there. We will here treat the methods listed above, in a fairly concise manner. We will also compare the numerical results of these more or less independent methods with each other, as well as with glacier and climate data.

2. Global averaging or series expansion of sea level trends

By calculating an average of the observed sea level trends s all over the world one might hope to eliminate the problem of vertical motions of the crust.

The idea is that, as a global average, the crust should be vertically stable. Denoting the rate of the global sea level rise by u we can simply write

$$u = \bar{s}$$

The first estimate of this kind was made by Gutenberg (1941), obtaining an average sea level rise of 1.1 mm/year (with a large standard error). Later Fairbridge (1961) obtained the same result.

The irregular geographical distribution of the sea level stations was also handled by Gutenberg. He grouped his stations into a smaller number of more equally spaced regions, calculated a mean trend for each region and then took the average of all regional trends as the final result. Similar procedures are still in use.

Nevertheless, no matter what averaging procedure is used, one drawback remains more or less unchanged: Sea level stations are inevitably mostly located along the coasts of the continents. This is a limitation in the sense that we cannot know for sure that the coasts, on an average, are vertically stable.

The problem of eliminating other sea level changes than the global rise has not been very much dealt with until more recently. The point here is to use only those sea level records that cover a sufficiently long time span, or at least give them a higher weight in the averaging procedure. Such attempts, using records covering most of the past century, have been made by Gornitz et al (1982), obtaining an average sea level rise of 1.2 mm/year, and Barnett (1983), obtaining 1.5 mm/year. It may be noticed that Barnett (1984), when limiting himself to the latter half of the original time span, obtained 2.3 mm/year, while e.g. Pirazzoli (1989) for a corresponding time period obtained about 0.6 mm/year (from mostly European and North American stations).

Ideally all sea level records used should be both long enough and cover one and the same time span. This requirement is very difficult to fulfil.

A new approach to the method of global averaging has been devised by Nakiboglu & Lambeck (1991). After forming regional mean trends for areas of 10° x 10° they expand this set of trends into a series of surface spherical harmonics,

$$s(\varphi,\lambda) = s_{00} + \sum_{n=1}^{K} \sum_{m=0}^{n} s_{nm}(\varphi,\lambda)$$

where φ and λ denote latitude and longitude, respectively. This yields a global long wave-length pattern of the sea level change with its possible regional deviations included. The zero degree term of the harmonic expansion represents the wanted global sea level rise:

$$u = s_{00}$$

The result of Nakiboglu & Lambeck amounts to

$$u = 1.15 \pm 0.38 \text{ mm/year}$$

The advantage of this approach is that regional effects, whether they be oceanographic or tectonic, are taken into account by the series expansion. This makes the solution less sensitive to regional irregularities.

There is one geophysical phenomenon that requires special attention when determining the global sea level rise, namely the postglacial rebound of Fennoscandia and Canada. This is so because the crustal uplift rate due to this phenomenon may amount to 10 times the rate of the global sea level rise, and to some extent it also affects stations outside the rebound areas. The traditional way of handling this problem has been to simply omit all stations in the rebound areas. However, a considerable part, in fact a majority, of all high quality long-term sea level stations in the world are concentrated to these areas, especially to Fennoscandia. Thus the omission of these sea level records imply a considerable loss of valuable information. During the last decade, therefore, one has tried to take advantage of these data, rather than to throw them away or let them go directly into a series expansion.

3. Postglacial rebound models and sea level trends

The phenomenon of postglacial rebound allows the determination of the viscosity of the Earth. The rebound after the melting of the large ice sheets over Canada and Fennoscandia is, in principle, characterized by an exponentially decaying uplift,

$$H(t) = H_0(1 - e^{-\alpha(\eta)t})$$

where H is the uplift since the deglaciation, H_0 is the total uplift = the depression at the time of the deglaciation, $\alpha(\eta)$ is a constant dependent on the viscosity η, and t is the time. The whole thing is observed as a corresponding decaying fall of sea level. Once you have a sufficient number of dated shore level observations to find the shape of the exponential curve you can determine the exponent and, thereby, the viscosity of the Earth's mantle. This was first made by Haskell (1935) from Fennoscandian shore-line data.

The idea behind using such a geophysical model of the postglacial rebound in the context of present global sea level rise is the following. The model will allow you to compute present uplift rates, $\dot{H}$, that do not include the global sea level rise. The observed sea level trends s, on the other hand, correspond to apparent uplift rates, $\dot{H}_a = -s$, that do include this sea level rise. Hence the discrepancy between the two kinds of uplift rates should reflect the rate of the global sea level rise:

$$u = \dot{H} - \dot{H}_a$$

It should be mentioned here that a simplified approach for finding $\dot{H}$, namely through linear extrapolation of the consecutive shore-line data, has sometimes been used. In this way Mörner (1973), Gornitz et al (1982) and Shennan & Woodworth (1992) found values for the sea level rise around 1.0 mm/year.

In reality, the geophysical modelling of the rebound is much more complicated than described above, for several reasons. The Earth is not purely viscous but rather viscoelastic. The mantle consists of at least two layers; different layers might have different viscosities. On the surface of this there is a solid lithosphere, the thickness of which needs to be taken into account. And on top of that we have the ice sheet itself, the thickness and shape of which also has to be modelled. (Its melting history is usually known from other sources).

Moreover, the sea level equation describing the height of the sea level in relation to the land does not only involve the uplift of the land. It also has to include the changing amount of sea water due to the melting of the ice sheets, as well as the changing load on the crust due to this redistribution of water. Furthermore, it has to include the change of the geoid caused by the change of the gravity field due to the redistribution of masses, both mantle, ice and water.

The first attempt to use such postglacial rebound models in determining present global sea level rise was made by Peltier & Tushingham (1989, 1991). Based on their viscosity and ice models they "corrected" all present sea level trends for postglacial rebound effects and then performed a global averaging, arriving at a global sea level rise of 2.4 mm/year. Subsequently Trupin & Wahr (1990) and Douglas (1991) have applied the same method, both arriving at 1.8 mm/year. We note that all these values are close to 2 mm/year, rather than the

1 mm/year obtained earlier. This lead the Intergovernmental Panel on Climate Change (IPCC) to somewhat revise their opinion on the global sea level rise accordingly; see Warrick & Oerlemans (1990) and Warrick et al (1996).

However, the modelling complications mentioned above require that a considerable amount of data is used in the modelling, covering the whole area as well as the whole time of the rebound. This is necessary to allow a reliable separation of the different quantities. Such a high resolution model has recently been introduced by Lambeck, Smither & Johnston (1998), based on a very large amount of geological shore-line data from Fennoscandia and its surroundings. This model has then been improved by Lambeck, Smither & Ekman (1998), using a consistent set of present apparent uplift rates from Fennoscandia calculated by Ekman (1996).

The improved model thus obtained includes ice thickness and shape, lithosphere thickness, upper mantle viscosity and lower mantle viscosity, and the analysis allows for a present climatic sea level rise. The outcome is a sea level rise of

$$u = 1.05 \pm 0.25 \text{ mm/year}$$

This value is valid for the 100-year-period 1892 - 1991, since this is the common time span for the Fennoscandian sea level trend calculations of Ekman (1996) used here. We note that the result is fully in line with the earlier one in Section 2 of Nakiboglu & Lambeck (1991), based on global averaging (series expansion). On the other hand it differs from the model results above, possibly because they all rely on more or less the same rebound model which seems to be based on insufficient data.

4. Vertical crustal motion recordings and sea level trends

An obvious method of separating vertical motions of the crust, such as postglacial rebound, from the global sea level rise would be to directly measure the crustal motion. This has hitherto not been quite possible for the simple reason that any vertical motion of the crust has been observable only in relation to the sea level itself. Nowadays, however, continuous three-dimensional satellite positioning, especially GPS, begins to offer a possibility, as noted by Johansson & Scherneck (1997) from Fennoscandian (Swedish) GPS data.

The idea behind this method is the following. Satellite positioning yields the absolute vertical motion or uplift of the crust. This is the same as the uplift relative to the reference ellipsoid, $\dot{h}$. Correcting this for the rise of the geoid, $\dot{N}$, which is not insignificant in rebound areas, we obtain the uplift of the crust relative to the geoid, $\dot{H} = \dot{h} - \dot{N}$. The sea level record yields, as before, the uplift relative to the sea level or the apparent uplift, $\dot{H}_a = -\dot{s}$. Then

$$u = \dot{h} - \dot{N} - \dot{H}_a$$

For the Fennoscandian rebound area, the geoid rise $\dot{N}$ here may be taken from Ekman (1998), based on Ekman & Mäkinen (1996), and the apparent crustal uplift $\dot{H}_a$ from Ekman (1996).

The absolute crustal uplift $\dot{h}$, however, is not yet easy to determine with sufficient accuracy. The most successful project in this respect is the one reported by Johansson et al (2000) based on continuous GPS recordings in Fennoscandia. The major problem to handle here seems to be systematic errors related to insufficient stability of the geodetic reference systems (ITRF) coupled to satellite orbits. Another problem is that the permanent GPS stations are normally not located close to the sea level stations. Clearly, however, the method as such should be considered as a promising future complement to the other methods for determining the sea level rise.

Another approach would be to make repeated absolute gravity measurements, instead of GPS, close to sea level stations. The gravity change rate thus obtained, $\dot{g}$, might be converted to the absolute uplift rate $\dot{h}$ through division by a constant k. Thus

$$u = \dot{g}/k - \dot{N} - \dot{H}_a$$

The quantity k equals $\dot{g}/\dot{h}$ as determined from the Fennoscandian land uplift gravity lines by Ekman & Mäkinen (1996), or any future improvement of that. The internal accuracy is somewhat less than for GPS but on the other hand the problems with the reference frame stability hardly enters here. However, k needs to be known with considerable accuracy which is a remaining problem.

5. Changes with time of very long sea level trends

A few sea level records in the world cover so many years that they allow a study of the change with time (acceleration) of the sea level rise. The one who drew attraction to this possibilty was Thorarinsson (1940). To start with, such records were used for estimating the sea level rise during the 1900s under the climatically based supposition that the rise during the 1800s was equal to zero, i.e. the change of the sea level trend between the two centuries was considered to be equal to the present sea level rise. In this way Lisitzin (1958, 1974), using the records of Brest and Swinemünde, obtained a sea level rise of 1.2

mm/year. Later Mörner (1973), using Amsterdam and partly Stockholm, and including geological aspects on vertical crustal motion there, obtained 1.1 mm/year.

Today the longest sea level records allow two things: a statistical analysis of the change of the sea level rise with time and, in combination with determinations of present sea level rise from other methods, a study of the sea level rise during earlier centuries. There exist three sea level records in the world that date back to the 18th century and, therefore, could be used for this purpose. They are:

Stockholm, commencing in 1774, the longest mean sea level series in the world and still in continuation (Ekman, 1988).

Amsterdam, commencing in 1700, the oldest mean sea level series in the world but discontinued in 1925 (van Veen, 1954; Spencer et al, 1988).

Liverpool, commencing in 1768, a high water series only but the longest of its kind (Woodworth, 1999).

Stockholm happens to be a most favourable station for studying long-term sea level changes. This is so for two reasons: Short-term (less than one month) sea level variations of non-Baltic origin, including tides, are filtered out at the Baltic entrance, and short-term variations of internal Baltic origin have a nodal line close to Stockholm (Samuelsson & Stigebrandt, 1996). Thus only the longer-term sea level changes show up here. Contrary to many other places in the world, this makes even weekly observations useful in calculating (monthly and) annual mean sea levels, on which the sea level trends are based. The sea level observations at Stockholm have, apart from some gaps during the first fifty years, mostly been made with the following frequencies: one reading per week 1774 - 1841; one reading per day 1842 - 1888; one reading per hour (from mareograph recordings) 1889 - present.

The Stockholm series has been analysed for general sea level rise by Ekman (1988, 1999). The annual mean sea levels were thereby corrected for the lunar nodal tide of period 18.6 years. The whole series of annual mean sea levels was divided into two parts of about 100 years each. By linear regression the rate of apparent postglacial uplift (uplift relative to the sea level) was found to be $\mathring{H}_a(t_1) = 4.92 \pm 0.23$ mm/year for the years 1774 – 1884 and $\mathring{H}_a(t_2) = 3.92 \pm 0.19$ mm/year for the years 1885 – 1984. (Adding the last decade would change the latter rate by less than 0.1 mm/year.) From this an increase in the general sea level rise amounting to

$$\Delta u = \dot{H}_a(t_1) - \dot{H}_a(t_2)$$

or

$$\Delta u = 1.01 \pm 0.30 \text{ mm/year}$$

is obtained. This is shown by a *t*-test to be highly statistically significant, at the 99.9 % level. The effect is clearly visible in Figure 1.

This change reflects the general rise (acceleration) in sea level due to the change from the colder climate at the end of the "Little Ice Age" during the first period to the milder climate during the second period, characterized by melting of glaciers. See further Section 6 and Figure 2, which reveals a general glacier retreat since the end of the 19th century.

One might argue here that there could be regional Baltic factors influencing the result above. The main such factors for most of the past century would be increasing salinity in the Baltic Sea (Kullenberg, 1981) and increasing south-westerly winds over the Baltic entrance (Ekman, 1998a). These two factors may be partly coupled, as winds affect the inflow of salt into the Baltic. However, their effects on sea level change should be very small, 0.1 mm/year or less (cf. Ekman & Mäkinen, 1996a, Samuelsson, 1996, and Ekman, 1998a). Moreover, they partly cancel each other since the effect of the winds is a higher sea level and that of the salinity a lower one.

Woodworth (1990, 1999a) has analysed the three oldest mean sea level records in the world, namely Amsterdam (1700 – 1925), Stockholm (1774 -) and Brest (1807 -), as well as the adjusted old high water level record of Liverpool (1768 -). Applying quadratic regression over their entire record lengths, he found a significant quadratic term in all four of them. Furthermore, this term is quite similar for the four sites, about 0.004 mm/year^2, corresponding to an acceleration of 0.008 mm/year^2 or 0.8 mm/year per century:

Amsterdam	0.8
Stockholm	0.9
Brest	0.9
Liverpool	0.7

This confirms that the Stockholm result above is not dependent on any specific conditions in the Baltic Sea but reflects a more general phenomenon.

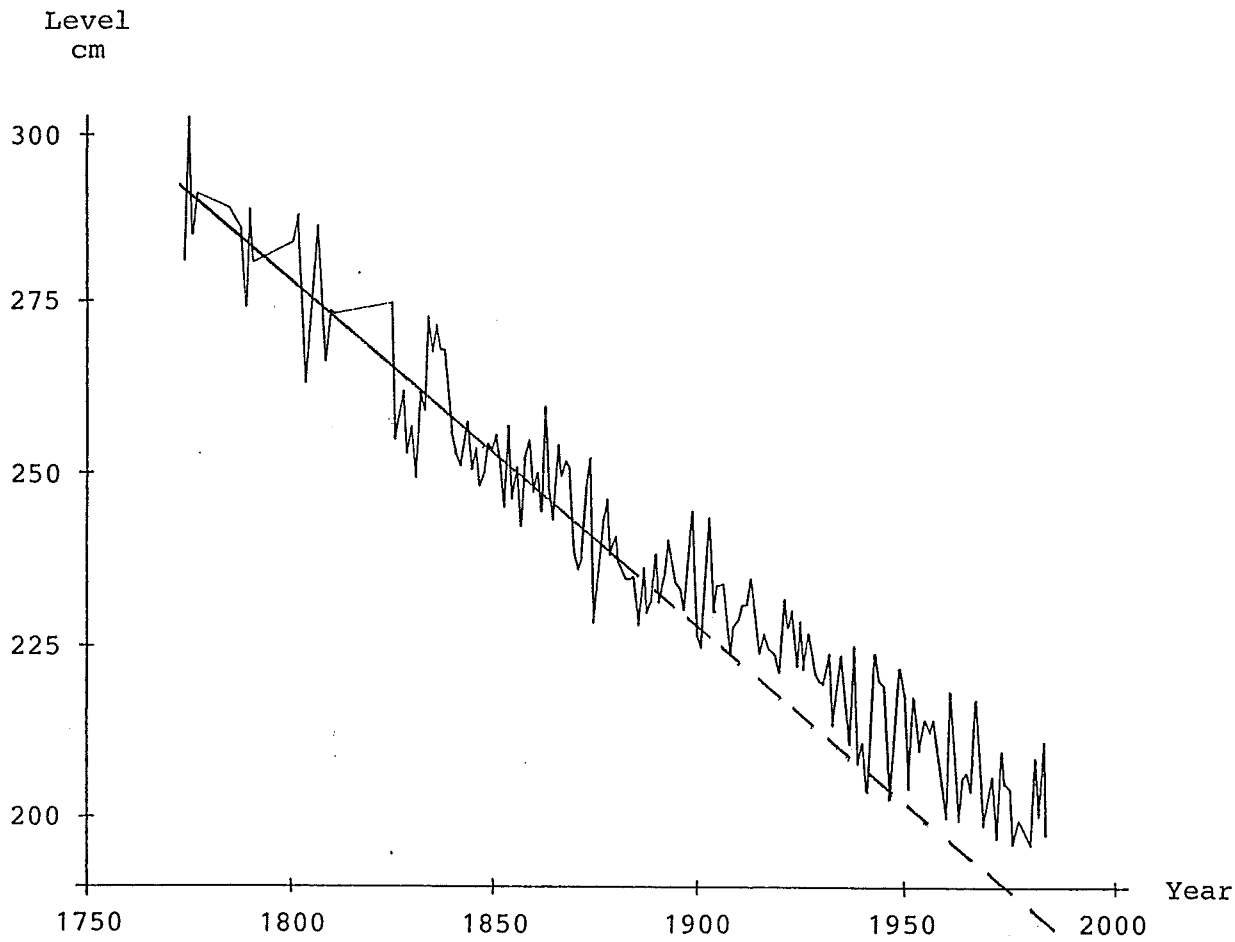

Figure 1. Annual mean sea levels at Stockholm since 1774 with the regression line for 1774 - 1884 and its extension into modern times.

Let us now return to the main result of Section 3 (which we remember also was in line with that of Section 2). There it was found that

$$u(1892\text{-}1991) = 1.05 \pm 0.25 \text{ mm/year}$$

This agrees almost precisely with the result for Δu above, showing that sea level rise during the end of the Little Ice Age was probably close to zero. We obtain

$$u(1774\text{-}1884) = 0.0 \pm 0.4 \text{ mm/year}$$

This is in good accordance with the sparse knowledge from glaciers as seen in Figure 2.

6. Comparison with climate changes

The global sea level rise seems to originate about equally from melting of mountain glaciers in the northern hemisphere and from thermal expansion of ocean water; see Warrick et al (1996). It is estimated there that they each contribute with 0.4 mm/year of sea level rise, but the uncertainties associated with these numbers are large. Already Thorarinsson (1940) in his pioneering study estimated that present glacier melting would account for about 0.5 mm/year of sea level rise. Later on similar results have been obtained by Meier (1984) and Zuo & Oerlemans (1997).

For some glaciers there exist data on their changes of length through the last few hundred years. These data reveal that most glaciers probably remained stable until the end of the 19th century when they started melting. A good illustration of this is given by Figure 2, taken from Warrick & Oerlemans (1990). As mentioned, the results in Section 5 for the sea level rise during different centuries are in good accordance with this knowledge of glacier retreat. This also implies that values of the sea level rise for the recently past century approaching 2 mm/year (quoted in Section 3) are unlikely, since that would indicate a sea level rise of 1 mm/year even during the Little Ice Age. This seems hardly possible from a climatic point of view.

The northern hemisphere climate changes during the last millenium are reasonably well known through three independent methods. These are the use of historical European documents, studies of tree rings from America, and investigation of oxygen isotopes in annual layers of the Greeland ice sheet. Combining these data sources, Hammer et al (1980) have established a northern hemisphere climate (temperature) curve. There, the climate changes during the time of the historical sea level records above can be seen to be

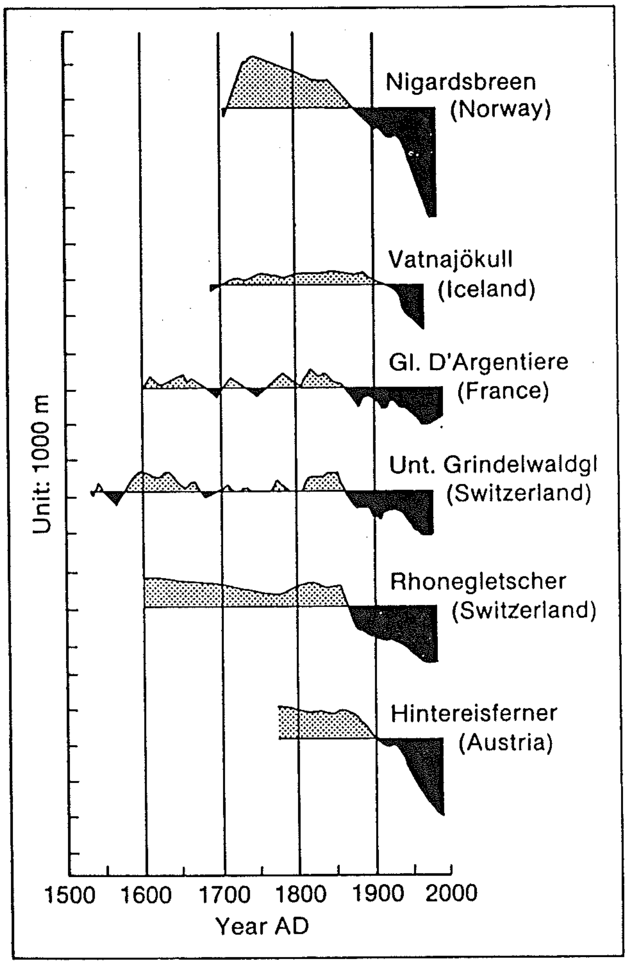

Figure 2. Changes of glaciers as measured by their length (Warrick & Oerlemans, 1990).

among the greatest ones since 800 A.D. (at least). This is even more emphasized in the recent climate curve by Mann et al (1999). Consequently, this should hold also for corresponding changes of sea level during the same period. Utilizing the results of Section 5 we may infer that global sea level changes due to northern hemisphere climate variations since 800 A.D. have probably always kept within - 1.5 and + 1.5 mm/year, with an average fairly close to zero.

Turning to the future, there is an understandable wish to identify a possible accelerated sea level rise due to the greenhouse effect (cf. Woodworth, 1990). The role of this effect in the climate warming of the past century is still controversial, but there are clear indications that the greenhouse effect has contributed to the warming during the last few decades (Tett at al, 1999; Karlén et al, 1999). We should point out here that it is very difficult to identify a sea level acceleration (see also Douglas, 1992) without the accelerated rise having gone on for quite a long time; at least this is so for the Stockholm record in the Baltic Sea. The main reason for this is the fact that during a shorter time interval, say one or a few decades, an apparent acceleration (or retardation) might very well be caused by anomalous winter wind conditions over the Baltic entrance; see Ekman (1997, 1998a). A better geophysical modelling of such phenomena might improve the situation somewhat.

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