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Global sea-level changes and their measurement

P.A. Pirazzoli

CNRS, Laboratoire de Géographie Physique, 1 Place Aristide Briand, 92190 Meudon-Bellevue, France

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ABSTRACT

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Field sea-level data are generally only of local value, being affected by a complex of local, regional and global processes which operate on different time and space scales. Averages or compilations of local sea-level data may lead therefore to misleading estimations of global sea-level changes and be biased towards predominant factors which are active in the study areas at the time scale considered. Approximate estimations of global sea-level changes can be attempted using proxy data, such as variations of oxygen isotope ratios in oceanic core sediments. These suggest an oscillatory pattern during the last 3 Ma, with glacial–interglacial cycles occurring with a periodicity of about 100 ka and sea-level fluctuations of the order of 100 m. Chronology in oceanic cores is calibrated with geomagnetic reversals, biostratigraphy and the identification of certain peaks in the isotopic curve, which allow a comparison with astronomical cycles. The amplitude of sea-level oscillations is refined through correlation with sequences of Quaternary marine terraces in uplifting areas. Holocene sea-level data show a great vertical dispersion, caused by the superposition of eustatic, isostatic, tectonic and other factors. Similar biases probably exist also for estimations of global sea-level change during longer (several Ma) and shorter (1–100 yr) geological time periods. This is true for tide-gauge data: only 13% of the stations with long enough records indicate a rise between 1.0 and 1.5 mm/yr (which corresponds to values often assumed as "eustatic"), whereas interpretations of global sea-level rise deduced from tide-gauge records diverge appreciably, with estimates for the past century ranging from 0.5 to 2.4 mm/yr. It is hoped that data from new altimeter satellites (ERS-1, TOPEX/POSEIDON), combined with Global Positioning System geodesy data, will soon clarify this uncertain situation.

Introduction

Sea-level variations correspond to a complex of local, regional and global processes which interact at various temporal, areal and vertical scales. Among the major causes of global sea-level changes, large scale geotectonic effects prevail over the long term (several Ma), climatic and isostatic effects over the mean term (10–100 ka), whereas over the short term (1–100 yr) a loud background noise (due to meteorological, hydrological, neo- and volcano-tectonic and anthropic factors) usually prevents accurate estimations.

Assessments of global sea-level changes have been attempted at varying time scales, by various disciplines and methods. Most of these assessments are based on simplified assumptions, owing to the methods used. In this paper, methods employed at certain time scales are reviewed, their reliability discussed and their accuracy estimated.

Over the very long term

The continued presence of life on the Earth's surface during the last 3.5 Ga, as proved by microfossils and stromatolites observed in many rock sequences going back to this time (Schopf, 1983), is evidence of a continued presence of water and oceans throughout this period. However, though it is known to be unlikely that the oceans evaporated completely (Kasting, 1989), great global changes in sea level did probably occur. According to a review by Emery and Aubrey (1991), during each of the five complete orogenic cycles which have occurred between the late Archaean (2.7 Ga) and the late Paleozoic (470 Ma), sea-level change may have ranged by about 500 m.

During the past 250 m.y. (application of seismic stratigraphy)

Seismic stratigraphy is a technique which has often been used to deduce sea-level fluctuations. It is based on the identification of regional surfaces of erosion or non deposition, called unconformities or sequence boundaries, which form during downward shifting of coastal onlap connected to relative sea-level fall (Vail et al., 1977).

By applying this technique, various sea-level curves have been proposed. Among the recent versions, the most complete curve is probably that proposed by Haq et al. (1987) (Fig. 1) to summarize global sea-level changes since 250 Ma. This curve was deduced from a compilation of stratigraphic sequences mostly from Europe and North America and shows changes of over 400 m in amplitude, between a peak more than 250 m above the present sea level in the Cretaceous and minimum levels at least 150 m below present during certain Plio-Quaternary glaciation periods. In comparison, the average sea-level rise, which would be induced by the complete melting of all land-based ice sheets and glaciers of the Earth, can be estimated at about 65 m at the present time and probably not much more than 200 m during maximum glacial times. A global sea-level change of the order of 400 m would therefore be impossible without major changes in the shape of the oceanic container, i.e. without important tectonic deformation (tectono-eustasy).

As plate tectonics have upset the shape of the oceans completely during the last 250 Ma, the occurrence of important tectono-eustatic variations appears very likely. Nevertheless the claim by Haq et al. (1987) that the curve of Fig. 1 represents "eustatic" (i.e. global) sea-level variations is hardly convincing. Existing compilations of relative sea-level data clearly show that no

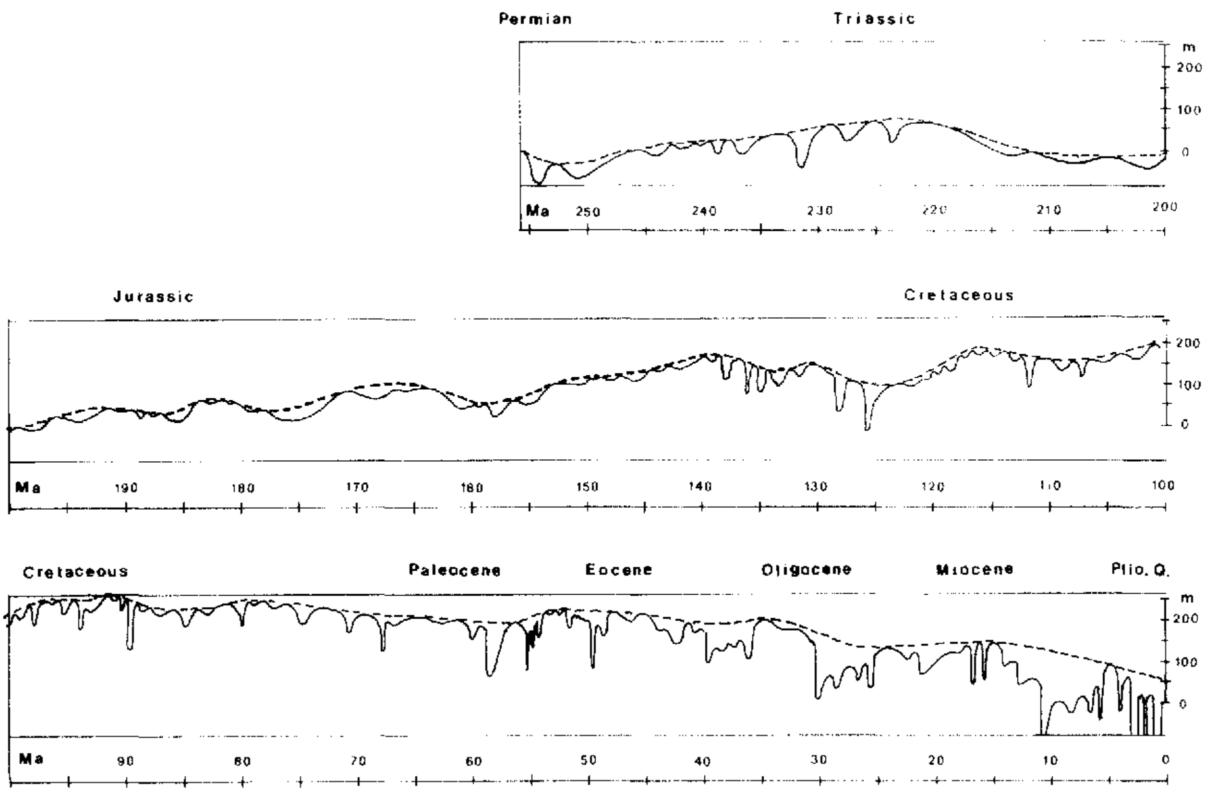

Fig. 1. Changes of "eustatic" sea level during the past 250 Ma, according to Haq et al. (1987) (adapted).

coastal region of the Earth can be considered as having been vertically stable during the past 10,000 years or even during the last century (see below). With greater reason the assumption of vertical stability during the past 250 Ma is illusory.

Apart from the obvious criticism that details of the data supporting the curve of Fig. 1 have not yet been published, several recent articles have focused on limitations of eustatic curves derived from seismic stratigraphy, discussing interpretation issues such as the origin and chronostratigraphic significance of seismic reflections, the precision with which unconformities can be identified and calibrated, the influence of tectonics, the significance and quantification of onlap, and regional bias (Christie-Blick et al., 1990).

Major geoidal changes (the amount of which is however unknown) are likely during the last 250 Ma to have affected the areas investigated by Haq et al. (1987) (Mörner, 1980a,b), especially on both sides of the Atlantic Ocean, which did not exist 250 Ma ago. As noted by Emery and Aubrey (1991, p. 59) stratigraphic relationships record changes in global sea levels, but also regional land movements and "averaging of stratigraphic data around the globe to remove these regional biases is certain to leave a residual bias on proposed sea-level curves". As an absolute reference datum is missing, "amplitudes of eustatic fluctuations cannot be inferred from seismic stratigraphic data alone because coastal aggradation is primarily a result of basin subsidence, not sea-level rise; and downward shifts in onlap reflect only the rate of sea-level fall relative to the rate of basin subsidence ... Which ever method is used to gauge changes in coastal onlap, the large component of subsidence cannot easily be removed to derive the smaller eustatic signal" (Christie-Blick et al., 1990, p. 136). As with most field sea-level data, stratigraphic data can therefore only provide information on local relative sea-level movements. Correlation may eventually give very useful information about the timing of sea-level fluctuations, on the time scale of thousands/millions of years, but little about global magnitudes. Lastly, relative sea-level history may eventually have been similar for passive margins on both sides of the Atlantic Ocean during certain periods, but extrapolation of this information to the global ocean would be arbitrary and unsound.

During the past 3 million years (oxygen isotopes records)

It is generally agreed that the oxygen isotope record which is obtained by analysing foraminifera in deep-sea sediment cores, can give a rough approximation of global ice volume (and therefore of global sea-level changes), because the isotopic composition of an ice sheet is generally lighter than that of the ocean water. Shackleton and Opdyke (1973) suggested the use of a change of about 10 m in the global sealevel for a 0.1‰ change in the isotopic record. Fairbanks and Matthews (1978) obtained a field calibration by measuring the isotopic composition of corals from raised terraces in Barbados, obtaining an estimate of 0.11‰ for a change of 10 m. These are only first approximations, however and Shackleton (1987) has summarized various sources of uncertainty: variations in the temperature of the ocean water, bioturbation, variations of the average isotopic composition of the former ice sheets in their size and their latitudinal position, etc. The accuracy of sea-level estimations from oxygen isotope records depends therefore on the validity of the assumptions made and could correspond, at best, to the precision of $\delta^{18}O$ measurements, i.e. to 0.1‰ or to ± 10 m in equivalent sea-level.

The chronology in oceanic cores is calibrated with geomagnetic reversals, biostratigraphy and the identification of certain peaks in the isotopic curve (e.g. stage 5e), which allow a comparison with astronomic periodicities. Recently, Shackleton et al. (1990) revised the Lower Pleistocene timescale on the basis of astronomical periodicities, shifting the last few reversals of the Earth's magnetic field by about 5–7% (Fig. 2).

According to Ruddiman and Raymo (1988), the initiation of moderate-sized ice sheets in the Northern Hemisphere occurred at 2.40 Ma, possibly preceded by small increases in ice volume at c. 2.55 Ma and by a longer-term high-latitude cooling that began about 0.75 Ma earlier. The history of ice-volume changes from 3.15 to 2.4 Ma is however still controversial, though all the evidence is consistent with a progressive, but oscillatory, deterioration of the northern hemisphere climate from 3.15 Ma (or earlier) until glaciations of substantial scale which began abruptly at 2.40 Ma. After 0.9 Ma, changes in $\delta^{18}O$ increased in amplitude by c. 50%, suggesting that ice sheets grew to considerably larger maximum volumes. Climatic oscillations show the predominance of the 41 ka cycle of orbital obliquity between 2.4 and 0.8–0.7 Ma, whereas during the mid-Pleistocene (0.9–0.4 Ma) a response at or near the 100 ka period of orbital eccentricity gradually increased. During the past 0.45 Ma the 100 ka period is predominant, with however numerous superimposed glacial advances at periods of 41 ka and 23 ka (Ruddiman and Raymo, 1988; Shackleton et al., 1988).

Careful comparisons made by Shackleton (1987), between the most detailed records of sea level over the last glacial cycle and high quality benthonic and planktonic isotope records deduced from 11 oceanic cores, have shown that during interglacial isotopic Stages 7, 13, 15, 17 and 19, when Holocene oxygen isotope values were not attained, the sea may not have reached its present level; either some northern hemisphere ice must have remained or ocean deep waters must have been colder than they are today. On the other hand the extremes of Stages 1, 5e, 9 and 11 are all so similar that it cannot be stated confidently that any one of these interglacials reached a significantly different level than another. Prior to Stage 11, probably the warmest interglacial in the last million years, it was only with Stage 25, at about 0.95 Ma, that there was another interglacial isotopically similar to the Holocene. Shackleton (1987) concluded that it is unlikely that sea level was glacio-eustatically higher than present by more than a few metres, during any interglacial of the past 2.5 Ma.

During the last million years (raised marine terrace correlations)

The record of global sea-level change can be interpreted from the study of ancient shorelines that have been tectonically uplifted or, in a similar way, from the analysis of glacial moraines record in areas rapidly subsiding isostatically

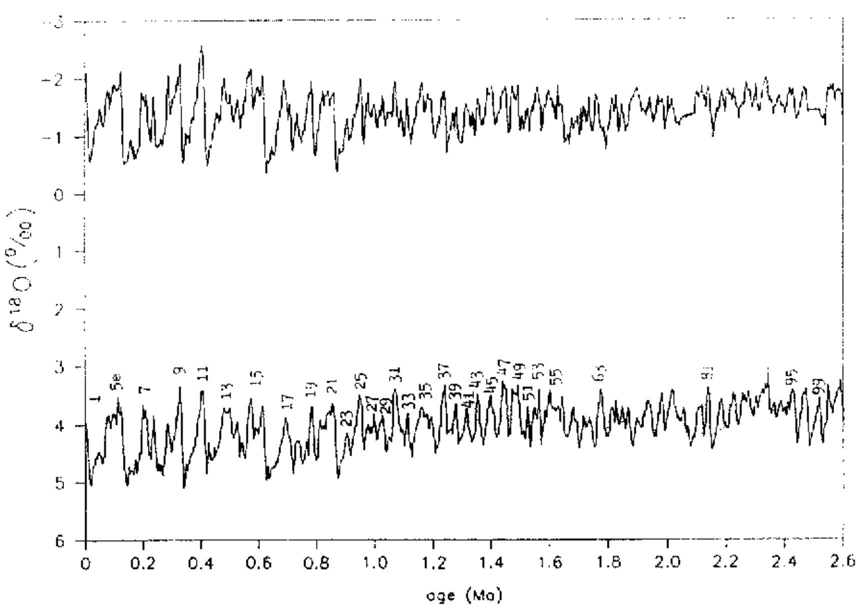

Fig. 2. Planktonic (above) and benthonic (below) oxygen isotope data for ODP Site 677 for the past 2.6 Ma, according to Shackleton et al. (1990). Labels of selected isotope stages have been added for orientation.

(Porter, 1979). In coral reef areas, uplifted terraces may be considered to be like a continuous tape recorder, each reef developing when the rising sea level overtakes the rising land; thus reef crests represent the peaks of each transgression. "The morphology and internal structure of each reef indicates the course of sea-level change relative to the rising land, so that sea level relative to stable ocean floor can be extracted for each section if the uplift rate for that section is known" (Chappell and Shackleton, 1986, p. 137). The assumptions usually made are: (1) that the eustatic sealevel position corresponding to at least one raised shoreline is known, and (2) that the uplift rate has remained essentially constant in each section. From these assumptions, sea level can be calculated for each dated reef crest. Chappell and Shackleton (1986) suggested the assumption that sea level at c. 125 ka (the 5e isotopic peak) was at +6 m; then, given the uplift rate U_t based on the height of the 5e reef in transect t, the sea-level position $S_{i,t}$ at the time the crest of reef i was formed in the same transect can be obtained from the equation:

$$S_{i,t} = H_{i,t} - U_t a_i \qquad (1)$$

where $H_{i,t}$ is the height of the crest of reef i on transect t and a_i is the age of reef i.

The lower levels which intervene between reef crests can be estimated in a similar way, using the heights of shallow marine and littoral deposits. However, less data are available to determine the low points, because the low sea-level deposits are generally buried by those of the subsequent transgression.

Bloom and Yonekura (1985, 1990) proposed a calculation procedure, slightly developed by Kikuchi (1988), in which the assumption of a constant rate of dislocation becomes unnecessary when a sequence of dated emerged shorelines is found at different heights on several transects; the heights of the intermediate shorelines can be plotted against the height of the highest one in each sequence; the regressions of the intermediate shoreline positions yield a set of equations:

$$H_{i,t} = r_i H_{m,t} + b_i \qquad (2)$$

where $H_{m,t}$ is the height of the highest shoreline in the sequence of transect t, $H_{i,t}$ is the height of an intermediate shoreline i on the same transect, r_i is the regression coefficient and b_i is the intercept. By assuming the position of sea level at the time of formation of the highest shoreline, the initial heights of the intermediate shorelines can be calculated by substitution in Eq. (2). A similar procedure can give the initial level of any shoreline in a sequence of submerged features, if their depths are plotted against the depth of the deepest shoreline in the same sequence.

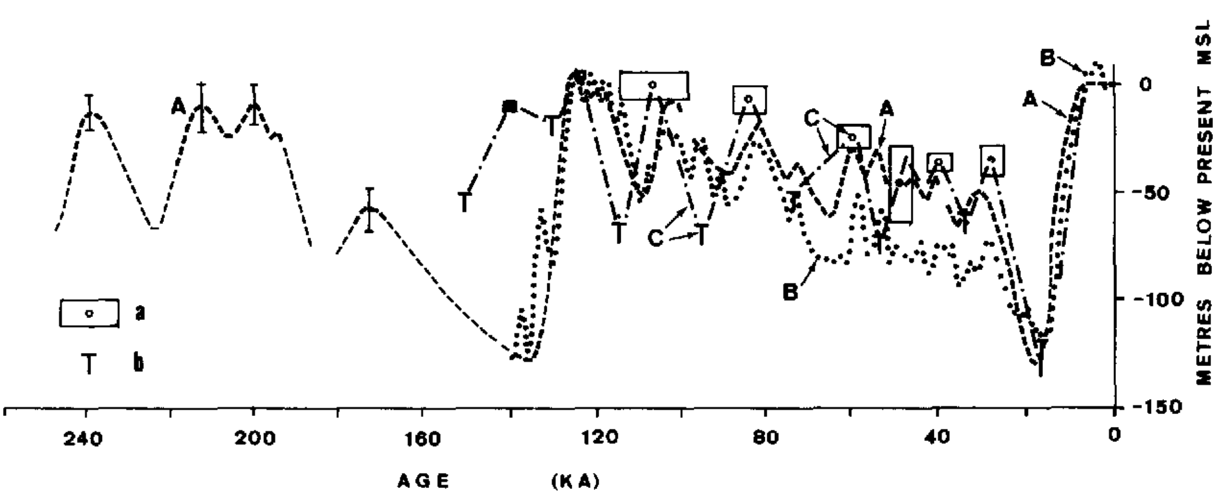

Fig. 3. Eustatic sea-level curves for the past 250 ka, according to Chappell and Shackleton (1986) (A), Shackleton (1987) (B) and Bloom and Yonekura (1985) (C). Boxes indicate probable errors of age and height of sea-level maxima (a) and minima (b) related to curve C. Vertical bars are uncertainty limits of altitude related to curve A.

Aragonitic corals from Pleistocene reef crests can be dated as far as 50 ka with the radiocarbon dating method and as far as about 350 ka with the U-series method. For higher and older terraces, the Electron Spin Resonance method made it possible to confirm the validity of correlations between raised marine terraces and isotope chronology back to isotope stage 15 (c. 600 ka) in Barbados (Radtke et al., 1988) and to stage 29 (c. 1 Ma) in Sumba Island, Indonesia (Pirazzoli et al., 1991, 1993). Similar correlations can also be made in volcanic regions where a tephrochronology is available, in New Zealand (Pillans, 1983) or Japan (Machida, 1975), or attempted by using amino-acid racemization methods (Muhs, 1983; Bowen et al., 1985; Ortlieb, 1987).

Examples of global sea-level curves deduced from sequences of marine terraces and oxygen isotopes data are given in Fig. 3. Curve *A* was obtained by Chappell and Shackleton (1986) from correlation between the coral terraces of Huon Peninsula and the ^{18}O record from core V19-30. Curve *B* was estimated by Shackleton (1987) using planktonic and benthonic data from various oceanic cores. The difference between curves *A* and *B* during the last glacial stage has been ascribed by Shackleton (1987) to a "step-like cooling in the deep water [which] occurred about 115 ka B.P. that was reversed about 11 ka B.P.", thus affecting the isotopic composition of benthonic foraminifera. Lastly, curve *C* is an estimate of the interstadial high sea levels of the last 105 ka deduced by Bloom and Yonekura (1985) from the Huon Peninsula data using Eq. 2 and an assumed sea level at +6 m at 124 ka.

During the last interglacial period

At oxygen isotopic stage 5e, a spread in radiometric dates and the occurrence in some sites of a double terrace (e.g. at Huon Peninsula, Papua New Guinea (Bloom et al., 1974), in Atauro Island, Indonesia (Chappell and Veeh, 1978), in Alor Island, Indonesia (Hantoro et al., 1993), etc. have been interpreted as indicating a prolonged stage, possibly interrupted by a minor sea-level fluctuation (Fig. 3). However no evidence of stratigraphic break has yet been reported. On the other hand stable isotope measurements in both molluscs (Shackleton and Matthews, 1977) and corals (Fairbanks and Matthews, 1978) from Barbados reefs have suggested that the high sea-level stand during substage 5e was probably rather short. A difficulty arises, however, when considering that erosional or depositional geomorphological features dating from 5e are often very well developed. The position of sea level around 140 ka remains therefore controversial. Recently, Lambeck and Nakada (1992) have concluded, using models of glacio-hydro-isostatic rebound, that the ocean volumes would have reached their present value by or before 135 ka and that substage 5e lasted until at least 120 ka. It can be expected that recent improvements in dating methods and techniques, especially with the use of mass spectrometry (Edwards et al., 1987; Bard et al., 1990) will soon help to clarify this point unambigously. Preliminary results (Bard et al., 1992) already seem to confirm an age range between 133 (or 135) ka and 120 ka, though doubts subsist about the real accuracy of the age determinations.

During the last hemicycle (local shorelines data)

From 1974 to 1982, when the International Union of Geological Sciences (IUGS) and Unesco sponsored the IGCP Project 61 "Sea level changes during the last hemicycle", under the leadership of A.L. Bloom, the main objective was the "establishment of a graph of the trend of mean sea-level during the last deglaciation and up to the present time; this graph will be an expression of the changing hydrologic balance between ice and water in response to climatic changes" [Geological Correlation, no. 2 (1974) to no. 11 (1983)]. However, it was clear from the first compilations of worldwide sea-level data (Pirazzoli, 1976; Bloom, 1977), later confirmed by more complete work (Newman et al., 1989; Pirazzoli, 1991), that local sea-level histories varied considerably around the world. Though part of this variability has been explained by global isostatic models, especially in areas near former ice sheets (Clark et al., 1978; Lambeck, 1990a,b; Tushingham and Peltier, 1991), significant deviations between model predictions and field data can still be found in many areas (Pirazzoli, 1991) and this confirms the fact that the relative sea level is affected everywhere by a complex of local, regional and global processes which operate on different time and space scales; it also suggests that no doubt global isostatic models attempt to take into account only a few of these processes, generally using assumptions which, in spite of their growing mathematical complexity, remain necessarily simple in comparison to the complexity of the earth's system.

The most important conclusion which could be obtained from IGCP Project 61, confirmed by models and by results of the subsequent IGCP Project 200 "Sea-level correlation and applications" (led by the present writer from 1983 to 1987), was that the determination of a single sea-level curve of global applicability is an illusory task.

As no stable area has been recognized anywhere in the world, similar conclusions can easily be extended to earlier periods of geological history (see above) as well as to tide-gauge records (see below). On the other hand, local shoreline data are essential to assess rates of vertical movement on a local scale and this often has important tectonic and geodynamic implications.

Global sea-level changes during the last hemicycle can be estimated, as during the past 3 Ma, using proxy data, such as oxygen isotope records or evaluations of changes in ice-sheet masses, or with assumptions on tectonic rates of vertical displacement. Anyway, very few sea-level field data exist on the shoreline position during the last glacial maximum; this position has been estimated at -175 m after 20,000 yr B.P., from dredging of shallow water reef coral and of other material suggesting a littoral origin in the Arafura Sea (Jongsma, 1970), and at -121 ± 5 m 17,000 yr B.P., from *Acropora palmata* corals drilled off south Barbados, assuming a continuous local uplift trend of 0.34 mm/yr (Fairbanks, 1989).

During the last century (tide gauges)

Though the establishment of the first tide marks dates from about three centuries ago, it was not until the second half of last century that major harbours started to be equipped with tide-gauge stations. At the beginning of this century, over one hundred and twenty stations were working regularly, mostly in Europe (especially along the Baltic coasts) and in a few harbours in other continents. Many other stations were installed during the first half of this century.

Very few gauges are sited on the open coast where a record would be most desirable for scientific work. Generally the gauge is associated with the operation of a major port and consequently is often to be found in a river or estuary. Recently, many new stations were installed for scientific purposes in the framework of the Global Sea-Level Observing System (GLOSS).

Monthly and annual mean values of sea level of over 1243 tide-gauge stations in the world were held in January 1987 in the databank of the Permanent Service for Mean Sea Level (c/o Bidston Observatory, Birkenhead, Merseyside L43 7RA, U.K.) (Pugh et al., 1987). Many additional data are also available for analysis (Spencer et al., 1988).

Systematic updating includes each year records from potentially about one thousand instruments which are presently in service: as many as 3000 new station-years values of mean sea level were entered in 1989 and over 2600 in 1990 (Woodworth, 1990b). Most of these records are however relatively brief and their geographical distribution very uneven (Fig. 4).

Following Gutenberg (1941), who was the first to publish an estimation of the recent eustatic variations in the world from tide-gauge records, various authors have attempted a similar evaluation (Table 1). Studies 1–15 were commented on in a previous review (Pirazzoli, 1989b) and also analysed by Emery and Aubrey (1991).

As concerns later estimates, Stewart (1989) made a critical examination of the results provided by Barnett (1984) and concluded, in agreement with Pirazzoli (1986), that relative sea-level rise is better explained by crustal movements than by eustatic sea-level rise.

Trupin and Wahr (1990), after correcting for the effects of postglacial rebound on individual station records, using the Peltier (1986) models for North America and northern Europe and data by Wagner and McAdoo (1986) for the remainder of the globe, estimated a global sea-level rise of between 1.1 and 1.9 mm/yr, with a "preferred value" near 1.75 mm/yr. Details of the stations considered and calculations made are not specified however.

Douglas (1991) considered only 21 stations (10 of which from the U.S.A., 18 from the North Atlantic and nearby Mediterranean areas) and found, after correction for the effects of post-glacial rebound using the ICE-3G model proposed by Tushingham and Peltier (1991), a "global" sea-level rise of 1.8 ± 0.1 mm/yr.

Three main approaches can be deduced from the studies summarized in Table 1. The first approach, common to most early studies (1–11) consisted mainly in using various kinds of averaging methods applied to uncorrected data, after exclusion however of record series considered doubtful or coming from areas regarded as uplifting or tectonic, but not excluding wide areas which are now known to be of moderate subsidence. This attempt, which obtained average values of sea-level rise ranging mainly from 1.0 mm/yr to 1.5 mm/yr, is considered to be biased towards a rise value greater than the eustatic one (Pirazzoli, 1986).

A second approach consisted in "correcting"

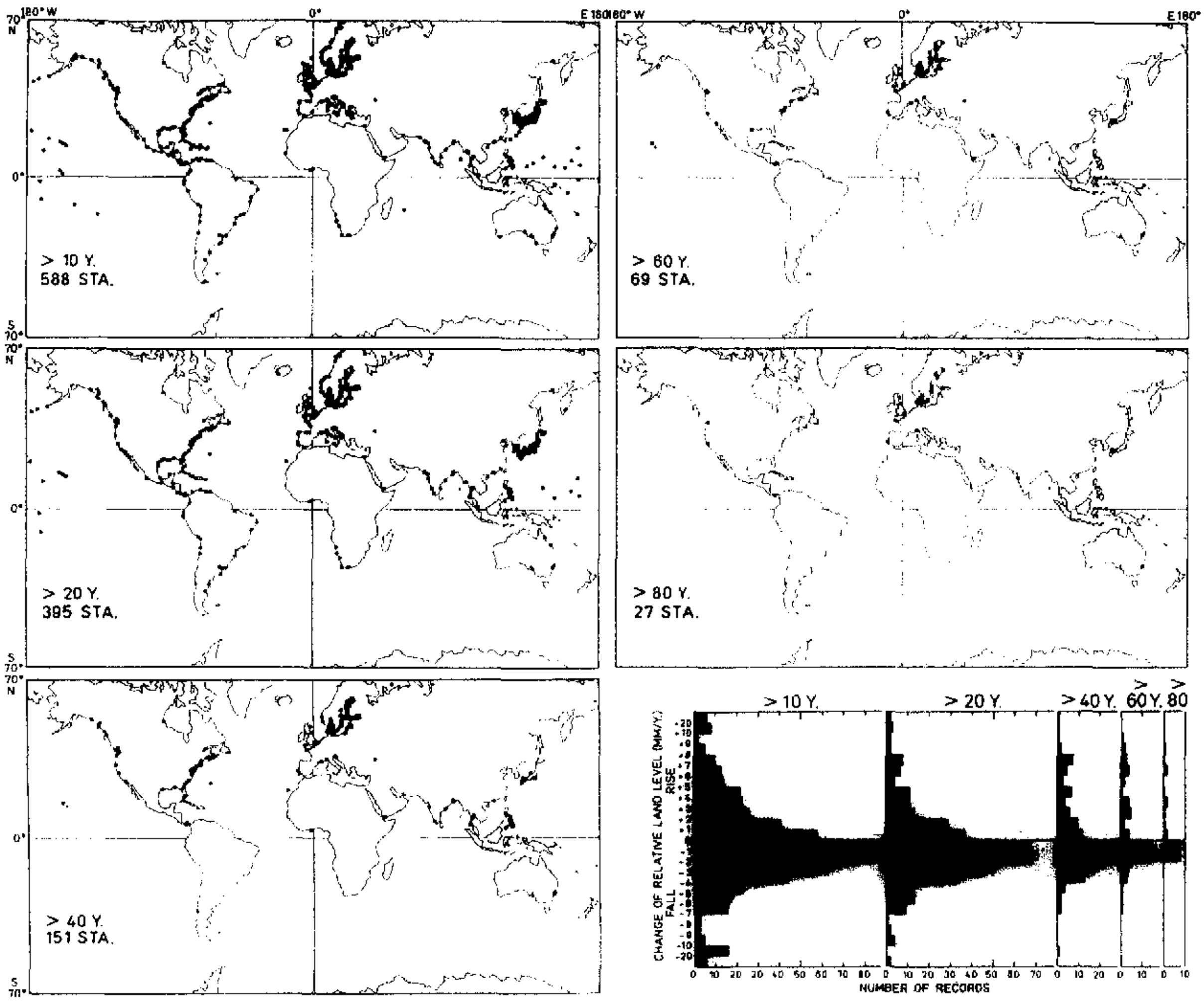

Fig. 4. World distribution of reliable tide-gauge records for periods of 10–80 yr, according to Emery and Aubrey (1991). Inset shows histograms for changes of relative land levels for each of the five subsets of station lengths. (Copyright Springer-Verlag, 1991, with permission).

the primary data, using geological data (9, 13), climatic correlations (15) or geophysical isostatic models (14, 17–18). This second attempt provided more variable results, though with greater values of computed sea-level rise when stations from the east coast of North America (where the rate of relative sea-level rise is generally much faster than in other coastal areas of the Atlantic) are predominant in the sample of tide-gauge stations considered.

The third group of studies (12, 16, 19–20) did not succeed in assigning any reasonable estimate to the present rate of eustatic rise of sea level, despite the evident need for such an estimate and the effort to determine that rate. This was due to a loud background noise (owing to meteorological, hydrological, tectonic and anthropic factors) and to the limitations of the data set used, which prevented accurate global assessments.

Only 13% of the 229 stations with long enough records considered by Pirazzoli (1986) indicate a rise between 1.0 and 1.5 mm/yr, and only 17%, a rise between 1.5 and 2.4 mm/yr (corresponding to values most often assumed as "eustatic"). On the other hand, 21% of the stations show a relative sea-level rise above 2.4 mm/yr, 20.5% a rise between 0.1 and 1.0 mm/yr, 1% a relative stability, and 27.5% a drop in the relative sea level.

Emery and Aubrey (1991) grouped 517 tide-gauge records according to geological processes or human interactions that appear to have controlled the vertical movements of land. They found very wide ranges of variation in all groups (Fig. 5). Coasts affected by glacial loading and unloading exhibit a range of relative land movement between about +15 mm/yr and −7.5 mm/yr, with a median of −0.2 mm/yr; belts of plate subduction and rifting, a range between +15 mm/yr and −23 mm/yr, with a median of −1.5 mm/yr; tide-gauge records at sites of volcanic activity, a range between +10 mm/yr and −15 mm/yr, with a median of −1.0 mm/yr, etc. Lastly 36 tide-gauge stations selected from "stable" coasts, i.e. from areas (p. 155) "where recent geological activities have been so minor as to allow these coasts to become the best sites for

TABLE 1

Estimates of global (average) sea-level rise from tide-gauge records

Authors	Number of stations	Period of time considered	Average rate of sea-level rise (mm/yr)
1 Gutenberg, 1941	69	1807–1937	1.1
2 Polli, 1952	110	1871–1940	1.1
3 Cailleux, 1952	76	1885–1951	1.3
4 Valentin, 1952	253	1807–1947	1.1
5 Lisitzin, 1958	6	1807–1943	1.1
6 Fairbridge and Krebs, 1962	unspecified	1860–1960	1.2
7 Kalinin and Klige, 1978	126	1900–1964	1.5
8 Emery, 1980	247	1850–1978	3.0
9 Gornitz et al., 1982	193	1880–1980	1.2
10 Barnett, 1983	9	1903–1969	1.5
11 Barnett, 1984	152	1881–1980	1.4
		1930–1980	2.3
12 Pirazzoli, 1986	229	1807–1984	indeterminable
13 Gornitz and Lebedeff, 1987	130	1880–1982	0.9–1.2
14 Peltier and Tushingham, 1989	40	1920–1970	2.4
15 Pirazzoli, 1989a	58 (Europe)	1880–1980	0.52
16 Stewart, 1989	152	1881–1980	indeterminable
17 Trupin and Wahr, 1990	84 (N of 30°N)	1900–1979	1.75
18 Douglas, 1991	21	1880–1980	1.8
19 Emery and Aubrey, 1991	517	1807–1986	indeterminable
20 Gröger and Plag, 1993	854	1807–1992	indeterminable

estimating present eustatic changes of sea level", indicate a sufficiently wide range (between +1.1 mm/yr to −4.2 mm/yr, with a median of −2.6 mm/yr) to prevent convincing eustatic assessments. The median displacement rate of all 517 stations considered acceptable by Emery and Aubrey (1991) is a relative land level fall of 1.1 mm/yr. However, as subsidence is more frequent than uplift in coastal areas, for geological reasons (Pirazzoli, 1986; Stewart, 1989), this value can be interpreted as corresponding to an average rate of crustal movements and hydrodynamic changes, though it may include an eustatic component of unknown amount. That is why Emery and Aubrey concluded (1991, p. 160) "we must state that we are unable to assign any reasonable estimate to the present rate of eustatic rise of sea level" and that the eustatic "rise can be bracketed only as ranging between 0 and 3 mm/yr".

Nevertheless, all European tide-gauge stations which are not uplifting show an average rate of relative land level fall of 1.26 mm/yr (Pirazzoli, 1989a); since, for rheological reasons, a subsidence movement in at least part of this area must necessarily exist, to balance the uplift movements in Fennoscandia, it can be concluded that the recent eustatic rise in Europe must certainly be less than 1.26 mm/yr.

On a global scale, Gröger and Plag (1993) consider that the value of a rise in global sea-level is a surface integral, the result of which is determined mainly by the spatial distribution of the sampling points, and that the sampling presently available is insufficient. A similar conclusion is

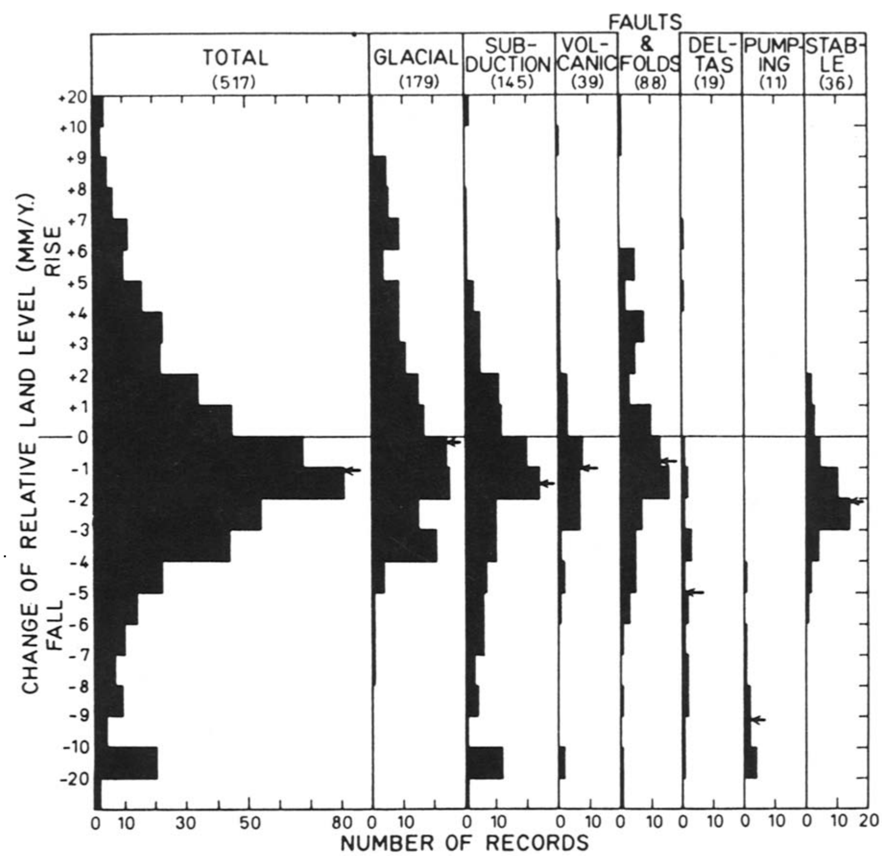

Fig. 5. Histograms showing distribution of directions and rates of vertical relative land-level change for 517 tide-gauge records, grouped according to geological processes that appear to have mainly controlled the vertical movements of land (from Emery and Aubrey, 1991). (Copyright Springer-Verlag, 1991, with permission).

reached by Dobrovolski (1992), for global temperature and sea-level changes. With a different approach, Mörner (1992) deduced from an analysis of the earth's LOD variations that the mean global sea-level rise during the last 150 years could have been at most 11 cm.

An important point on which general agreement is emerging, is that, in spite of increasing concentrations of "greenhouse" gases in the atmosphere, no evidence can yet be found for mean sea level rise accelerations (e.g. Woodworth, 1990a).

At the present day (altimeter satellites measurements)

During the last two decades, satellites have offered an unprecedented opportunity to obtain a global scale coverage of the ocean surface at regular time intervals. What is measured by satellites is the nadir H_1 distance from the satellite altimeter to the instantaneous surface of the ocean (Fig. 6). To establish sea-level changes, what has to be determined is H_2, whereas H_3 and H must be known accurately. H_2 includes all oceanographic effects, i.e. tide, currents and associated eddies, meteorological effects and steric effects. The accuracy of satellite altimetry depends on altitude measurement errors and radial orbit errors. As concerns geoid height, ocean dynamic topography and tides can also be considered as error sources.

The first satellite provided with an altimeter, Skylab, launched in 1973, had a range resolution of about ± 1 m. With the GEOS-3 satellite, resolution improved two years later to about 0.5 m. With SEASAT, in 1978, it reached 0.1 m (Pugh, 1987). GEOSAT, launched in 1985, which functioned successfully for almost 5 years of continuous operation, with a resolution similar to that of SEASAT, nearly achieved the full potential of this method (Sandwell, 1991).

The study of relative sea-level change in a given area is made possible only if a reference datum is chosen in that area. Since most vertical datums are referenced to mean sea level, since mean sea levels do not lie exactly on a equipotential surface due to the effects of currents and other oceanographic phenomena, and since local mean sea level diverges from the geoid by about ± 1 m, the vertical datums of the world are only consistent at the ± 1 m level.

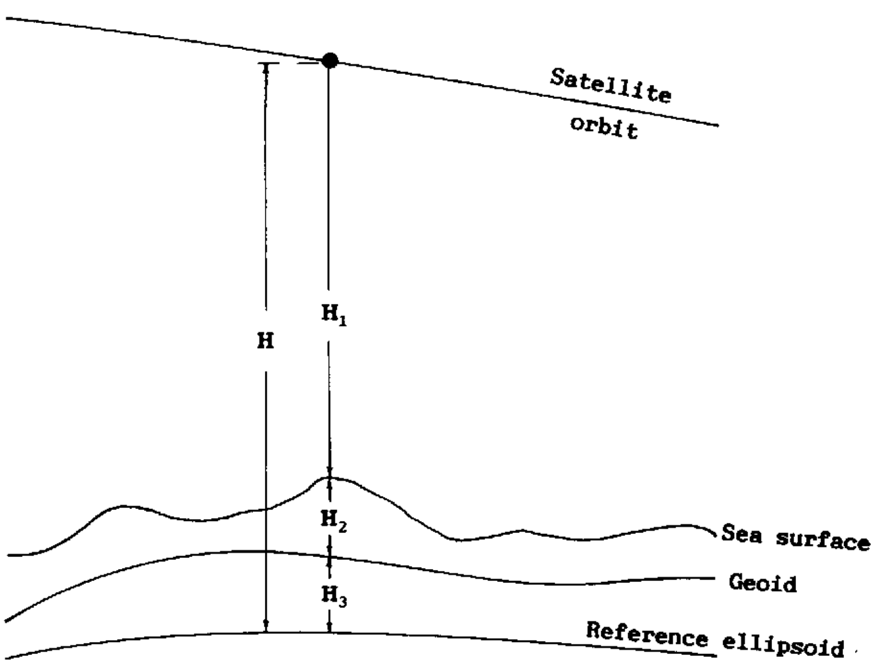

Fig. 6. Parameters involved in interpreting height H1 measured by a satellite altimeter above the sea surface.

In order to document present-day global sea-level changes, which may be of the millimetre per year order, datums should be correlated at least on a decade basis at a ± 1 cm level. This requires an order of magnitude improvement in our ability to characterize sea-level trends.

With the ERS1 and TOPEX/Poseidon satellites (the latter has been launched successfully on 10 August 1992), resolution is expected to be brought down to a few centimetres. According to preliminary results (Fu, 1992), the precision of the two altimetres have met the 2-cm (root-mean-square) specification. This, and a combination of satellite altimetry with Global Positioning System (GPS) geodesy and improved tide gauges, promise to provide the required 1-cm vertical accuracy in relating tide gauge datums to Very Long Baseline Interferometry (VLBI) and to tide-gauge datums in selected regions, e.g. in the framework of the Project "Space geodesy and global sea level" of the International Lithosphere Programme (Zerbini and Bilham, 1990).

Conclusions

The absence of vertically stable reference datums has until now prevented accurate estimations of global sea level change to be made, at practically all time scales.

Over the very long term, the range of sea-level change is estimated to be about 500 m, but the accuracy of such an evaluation is unknown. During the past 250 Ma, sea-level oscillations in the range of 400 m have been inferred from seismic stratigraphy (Fig. 1), but these are suspected of including tectonic components of unknown amounts.

For the past 3 Ma, oxygen isotope records provide very precise indications of eustatic sea-level changes, with an accuracy at best of ± 10 m. However, when correlations with well-studied coral terraces can be made for the late Pleistocene, an accuracy of the order of ± 10 m only seems realistic, with assumptions, during interglacial high sea levels, whereas uncertainties due to changes in deep ocean water temperature may be several times greater during glacial stadials.

In the course of the last hemicycle, geoidal changes of glacio-isostatic and hydro-isostatic origin have been greater than eustatic changes. Even in the last few thousand years, when eustatic changes are estimated to have been small or non-existent, a comparison between some 800 published local sea-level curves from around the world shows changes in the relative sea level greater than ± 5 m in 55% of the cases since 5000 yr B.P., and in 25% of the cases since 2000 yr B.P. (Pirazzoli, 1991). This of course prevents the use of local data for accurate assessment of global sea-level changes in the absence of accurate information on local isostatic movements or other tectonic trends.

Finally during the past century, interpretations of global sea-level rise deduced from tide gauge records have diverged appreciably. Though no author claims that sea level has been dropping, estimates of the rise during the past century range from 0.5 to 3.0 mm/yr. Some authors (including the present writer) even believe that the rate of this rise cannot be determined from tide-gauge records in a simple way.

In conclusion, in the absence of a stable datum, uncertainties seem to be of the same order as the changes in global sea level at most of the time scales considered. This proves that major progress, at least for present-day trends, can be expected in the establishment of absolute datums with narrow uncertainty margins, thanks to increasing improvement in satellite altimetry.

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