Temporal variation of oceanic spreading and crustal production rates during the last 180 My

Jean-Pascal Cogne\textsuperscript{a,\*}, Eric Humler\textsuperscript{b,1}

\textsuperscript{a}Laboratoire de Paléomagnétisme, UMR CNRS 7577, Institut de Physique du Globe de Paris, 4 place Jussieu, 75252 Paris cedex 05, France

\textsuperscript{b}Laboratoire des Géosciences Marines, UMR CNRS 7097, Institut de Physique du Globe de Paris, 4 place Jussieu, 75252 Paris cedex 05, France

Received 3 October 2003; received in revised form 8 August 2004; accepted 1 September 2004
Available online 7 October 2004
Editor: E.Bard

Abstract

We present a re-evaluation of seafloor spreading and generation rates, mainly based on a direct measurement of the remaining surfaces of oceanic crust and isochron lengths defined in the most recent isochron maps [J.Y. Royer, R.D. Müller, L.M. Gahagan, L.A. Lawyer, C.L. Mayes, D. Nürnberg, J.G. Sclater, A global isochron chart, Tech. Rep. 117, Austin, Univ. of Tex. Inst. for Geophys., 1992; R.D. Müller, W.R. Roest, J.Y. Royer, L.M. Gahagan, J.G. Sclater, Digital isochrons of the world’s ocean floor, J. Geophys. Res., 102 (1997), 3211–3214]. Our evaluation of the amount of oceanic crust per unit age \(\frac{dA}{dt}\) as a function of age, which can be expressed as \(\frac{dA}{dt} = C_0(1 - \frac{t}{t_m})\), is in fairly good agreement with previous determinations [J.G. Sclater, B. Parsons, C. Jaupart, Oceans and continents: similarities and differences in the mechanisms of heat loss, J. Geophys. Res., 86 (1981) 11,535–11,552; D.B. Rowley, Rate of plate creation and destruction: 180 Ma to present, Geol. Soc. Amer. Bull., 114 (2002) 927–933], with \(C_0 = 2.850 \pm 0.119 \) km\(^2\) year\(^{-1}\) and \(t_m = 180.2 \pm 9.7\) Ma. Dividing these \(\frac{dA}{dt}\) by the ridge lengths \(L\), defined as the isochron length at each epoch allowed us to compute the evolution of global half-spreaading rates. These have been roughly constant at 25.9 \(\pm 3.3\) mm year\(^{-1}\) for at least the last 150 Ma. We propose that the global seafloor surface generation rate is roughly constant as well, with a mean half-value of 1.298 \(\pm 0.284\) km\(^2\) year\(^{-1}\) and varying \(\pm 20\)% with time. This study corroborates the recent conclusion of Rowley [D.B. Rowley, Rate of plate creation and destruction: 180 Ma to present, Geol. Soc. Amer. Bull., 114 (2002) 927–933], of a constant generation rate since 180 Ma, and completely contradicts the commonly accepted idea of high seafloor spreading and surface generation rates during a large part of the Cretaceous. Combining the oceanic surface generation rates derived here with crustal thicknesses deduced from the chemical composition of old oceanic crusts and seismic measurements [E. Humler, C.H. Langmuir, V. Daux, Depth versus age: new perspectives from the chemical compositions of ancient crust, Earth Planet Sci. Lett., 173 (1999) 7–23], the magmatic flux at young (0–80 Ma) oceanic ridges appears to be about 18.1 \(\pm 3.4\) km\(^3\) year\(^{-1}\) and was possibly 15\% to 30\% higher during the Mesozoic.

\* Corresponding author. Tel.: +33 1 44 27 60 93.
E-mail addresses: cogne@ipgp.jussieu.fr (J.-P. Cogné), humler@ipgp.jussieu.fr (E. Humler).
1 Tel.: 33 1 44 27 50 88.

0012-821X/S - see front matter © 2004 Elsevier B.V. All rights reserved.
propose that mantle temperature variation provides an alternative mechanism to spreading rate for the Cretaceous highstand in sea-level and atmospheric CO₂ generation. © 2004 Elsevier B.V. All rights reserved.

Keywords: seafloor spreading rate; ridge length; magmatic flux; sea-level; atmospheric CO₂; Cretaceous; mantle temperature

1. Introduction

Most of the Earth’s volcanism occurs at mid-ocean ridges and forms the oceanic crust. Until recently, spreading rates and mean crustal thicknesses were difficult to evaluate and the amount of crust generated at ridges was poorly estimated. Menard [6], Deffreyes [7] and Dickinson and Luth [8] estimated the rates of basalt production at spreading centers at about 5–15 km³ year⁻¹ since the late Mesozoic. In the last 20 years, considerable advances in marine geosciences have allowed a better estimate of the production at oceanic ridges, which is now accepted to be about 20 km³ year⁻¹ (e.g. [9–11]). This value and its temporal evolution are important because they are used in various models of the Earth dynamics such as: heat flux balances [3,12], geochemical mass balance calculations (e.g. [9,10]), sea-level changes [13,14] and the CO₂ evolution of the atmosphere (e.g. [15]).

It is often assumed that the Cretaceous sea-level highstand (e.g.[16,17]) is the result of increased ridge volumes linked to rapid sea-floor spreading during that period, following the mechanism proposed by Hays and Pitman [13] (e.g. [18,19]). This idea of rapid mid-Cretaceous seafloor spreading, however, has been strongly debated, and is judged unnecessary, and perhaps wrong by Heller et al. [20], who believe that it is an artifact of poorly defined timescales. It should be noted, as Hardebeck and Anderson [21] did, that (1) the inferred high rates took place during the Cretaceous Long Normal Superchron (LNS), where there are no reversals to precisely determine the seafloor generation history; (2) the original derivation of these rates [14] was based mainly on reconstructions of the Pacific Ocean assuming symmetric oceans of which a large part has now been subducted; (3) ridge jumps were not taken into account, which can result in a large overestimation of seafloor spreading; (4) the influence of other oceans (so-called «Gondwana» oceans such as Atlantic and Indian oceans) was underestimated.

Rowley [4] recently reappraised seafloor spreading rates since 180 Ma, using the oceanic age grid proposed by Royer et al. [1] and Müller et al. [2]. Integrating the differential surface/age {dA/dt} distribution proposed by Sclater et al. [3] and Parsons [27], which is of the form {dA/dt}=C_o(1−t/t_m), where C_o is the present-day oceanic crust production in km² year⁻¹, and t_m the maximum age of the crust, the author proposed that plate production remained roughly constant since 180 Ma at 3.4 km² year⁻¹. This conclusion, however, is based on two main assumptions, which are (1) the destruction of oceanic floor at subduction zones is uniformly distributed with age, and (2) from present to 180 Ma, the oldest age of oceanic crust remained constant at 180 Ma. Both hypotheses may be debated, but their validity while commonly accepted is difficult to prove (e.g. Parsons [27], Rowley [4]).

For these reasons, we decided to revisit the seafloor production rates, using a different method: the direct measurement of currently visible seafloor surfaces encompassed between pairs of isochrons in each of the main oceanic basins. Our work follows the method of Sclater et al. [3], but uses the more recent, and accurate, database of Royer et al. [1] and Müller et al. [2]. Dividing surfaces by the isochron length, which are the ridge remnants which produced these surfaces, we thus obtain a figure of spreading rates on each of the main oceanic ridges, which we then average to evaluate the mean global spreading rate since 180 Ma. In a second and more speculative step, we compute the global seafloor surface generation rates since 180 Ma by applying our derived ridge spreading rates to hypothesized global ridge lengths including subducted ridges of the Pacific and Tethys oceans [22,23] (see Section 3.3). Finally, in Section 3.4, we estimate the temporal variation of oceanic crust production, in volume, in
2. Method

2.1. General

We have based all of the present study on the synthetic isochron database published by Royer et al. [1] and Müller et al. [2]. This database (which may be downloaded at ftp://ftp-sdt.univ-brest.fr/jyroyer/Agegrid/) reconstructs magnetic anomalies 5, 6, 13, 18, 21, 25, 31, 34, M0, M4, M10, M16, M21 and M25, calibrated using geomagnetic timescales of Cande and Kent [24] for anomalies younger than chron 34 (83 Ma), and of Gradstein et al. [25] for older periods. An error gridmap is also provided, computed as a function of (1) errors on ocean floor ages, (2) distances of each grid cell to the nearest magnetic anomaly, and (3) the gradient of the age grid. Except in the vicinity of large fracture zones, most part of this age map is estimated to be defined with an accuracy better than ±3 Ma.

We first computed the area of oceanic crust \{dA\} comprised between each successive pair of isochrons, in each of the main oceanic basins (Atlantic and Indian Oceans, Antarctic, Pacific, Nazca, and Cocos plates), using the PaleoMac application of Cogne [26] (Fig. 1). We imposed an error of ±3% on \{dA\} from the computations of the spherical shells defined by the isochron outlines.

Second, we estimate the area of ocean floor per unit age \{dA/dt\} as a function of age by dividing each \{dA\} by the time \{dt\} between each isochron pair (Fig. 2; Table 1, and Table A in the online appendix). A comparison with the estimates of Sclater et al. [3] will be discussed below. The errors on \{dt\} have been largely discussed in the literature, and may be the cause of significant variations in \{dA/dt\} (e.g. [2,14,19]). Following the error map proposed by Royer et al. [1] and Müller et al. [2], we assume an average constant error of ±1 Ma on each isochron, thus on each area boundary, resulting in a ±√2 Ma error on the \{dt\} parameter.

Finally, we calculated spreading rates by dividing the \{dA/dt\} values by the isochron length \{L\} (Table A in the online appendix) at each epoch. \(L\) is computed by summing up the angular distance between each pair of points defining the “productive” segments of each isochron (Fig. 1). We quantified the uncertainty on \{L\} by comparing the measurements of symmetric pairs of isochrons in both (symmetric) Atlantic and Indian basins. It appeared to vary between 5% and 10% of the total length. We therefore assumed a 7% error on these determinations. The \{dA/dL\} parameter (Fig. 3; Table A in the online appendix) obtained for each of the Atlantic, Indian, Pacific, Nazca, Cocos and Antarctic basins expresses the area of oceanic crust per unit age and unit ridge length as a function of age. This therefore represents an average spreading rate (in mm year\(^{-1}\)) for each basin. The error on these computations was obtained using the classical error propagations calculations on \{dA\}, \{dt\} and \{L\}.

2.2. Atlantic Ocean

Because the Atlantic Ocean is symmetric and bounded by passive margins, the \{dA\} parameters (Table A(a)) were simple to measure. We arbitrarily cut the Atlantic Ocean from the Indian Ocean at about

the light of the recently proposed temperature variations in the upper mantle [5].
Fig. 2. Area of the ocean floor per unit age \( \frac{dA}{dt} \) as a function of age for (a) Atlantic, (b) Indian, (c) Antarctic, (d) West Pacific, Nazca, and Cocos and (e) Global ocean floors. The straight dotted line in (e) is the best-fit line over points given by Eq. (1) with \( C_0=2.85 \text{ km}^2 \text{ year}^{-1} \) and \( t_m=180.2 \text{ Ma}. \)
10°E longitude, south of Africa. The ridge lengths \{L\} given in Table A(a) were evaluated by averaging the measure of symmetric pairs of isochrons from both sides of the Atlantic Ridge.

2.3. Indian Ocean

The Indian Ocean is a little bit more complicated. It is composed of symmetric crust segments around three main ridges: the Central Indian Ridge (CIR), the South East Indian Ridge (SEIR) and the South West Indian Ridge (SWIR). We calculated the ridge lengths \{L\} by averaging the values measured on symmetric isochron pairs from both sides of these ridges as in the Atlantic Ocean (Table A(b)). Similarly, we computed the spreading rates \{dA/dt/L\} using \{dA\} values from both sides, including the Antarctic part of this ocean, south of SEIR and SWIR. However, to compute the Global \{dA\} of Table 1, we had to exclude the Antarctic part of the Indian Ocean, to avoid adding this crust segment twice. The \{dA\} and \{dA/dt\} parameters of the Indian Ocean excluding the Antarctic part are given in Table A(c).

2.4. Antarctic plate

The surface \{dA\} and isochron lengths \{L\} of Antarctic plate are given in Table A(d). They allowed to compute spreading rates \{dA/dt/L\} which, compared to Atlantic and Indian Oceans, represent half-spreading rates. It should be noticed that the global data given in Table 1 do include \{dA\} from the Antarctic, but exclude ridge lengths \{L\} from this plate because they are already included in the Indian, Pacific and Nazca estimates (see below).

2.5. Pacific, Nazca and Cocos plates

The Pacific plate is bounded to the east by the East Pacific Rise (EPR), and to the south by the Pacific–Antarctic Rise. The Nazca plate is bounded by the Chile Rise, the EPR and the Cocos–Nazca Ridge, and the Cocos plate is bounded by the EPR and the Cocos–Nazca Ridge. The spreading rates \{dA/dt/L\} (Table A(e, f and g)) were determined independently within these three plates and represent half-spreading rates. However, the total ridge lengths given in Table 1 were obtained by summing the Pacific \{L\} with only the Cocos–Nazca and Chile Rise parts of the Nazca plate, which are given in Table A(h).

2.6. Summary

The Global surfaces \{dA\} of Table 1 were obtained by summing the Atlantic, Indian without Antarctic, Antarctic, Pacific, Nazca and Cocos surfaces quoted in Table A. The Global ridge lengths \{L\} (Table 1) were calculated by summing the Atlantic, Indian, Pacific, Chile and Cocos–Nazca \{L\} of Table A, excluding other Nazca and Cocos isochrons as well as Antarctic isochrons.

3. Results

3.1. Surfaces

The area of oceanic crust per unit time \{dA/dt\} as a function of time (Table 1, and Table A in the online
Fig. 3. Half-spreading rates on (a) Atlantic, (b) Indian, (c) Antarctic, (d) Pacific, Nazca, and Cocos ridges as a function of age. (e) Average half-spreading rates without (thin line) and with (bold line) the Pacific, Nazca and Cocos data. The bold dotted lines with dark grey areas are the averages computed for the 0–33.1, 33.1–120.4, and 120.4–147.7 Ma periods (see Section 3.2).
appendix) is illustrated in Fig. 2 for each of the main ocean basins. It is clear from this plot that a large variability exists between them. Although seafloor production of the Atlantic has remained fairly stable since the opening of the South Atlantic Ridge, the Indian ocean exhibits a jump in production between chron 21 and 31 (47.9 to 67.7 Ma), the period during which the India plate rapidly drifted to the north, before colliding with Eurasia. The Antarctic plate displays a regularly increasing spreading rate since chron M0, which is due to the fact that it is entirely bounded by ridges. The West Pacific plate exhibits a more sporadic production rate, with period of constant production between chron M21 and M0 (147.7 to 120.4 Ma), and between M0 and isochron 13 (120.4 to 33.1 Ma).

Early methods to evaluate the evolution of the (observable) total amount of oceanic crust per unit age \(\{dA/dt\}\) as a function of time \(\{t\}\) employed the relationship:

\[
\frac{dA}{dt} = C_o \left(1 - \frac{t}{t_m}\right)
\]

where \(C_o\) is the present-day oceanic crust production in km\(^2\) year\(^{-1}\), and \(t_m\) the maximum age of the crust (Sclater et al. [3] and Parsons [27]). They found roughly linear rate on global scale with \(C_o\) of 3.45 km\(^2\) year\(^{-1}\) and \(t_m\)=180 Ma, for a total measured surface of 310 \(\times\) 10\(^6\) km\(^2\). The more recent evaluations of these parameters by Rowley [4] give \(t_m\)=182 Ma and \(C_o=2.960\) km\(^2\) year\(^{-1}\).

Our results are similar to those one proposed by Sclater et al. [3] and Rowley [4]. However, the best fit line over our points defines a \(C_o\) parameter of 2.850 ± 0.119 km\(^2\) year\(^{-1}\) which is significantly lower than the value reported by Sclater et al. [3]. Our \(t_m\) is strictly identical (\(t_m=180.2\pm9.7\) Ma) to that of Sclater et al. [3] but the surface either measured (268 \(\times\) 10\(^6\) km\(^2\)), or obtained by integrating the above equation (1/\(2C_o t_m=257\times10^6\) km\(^2\)) is significantly (~15%) lower than their value. However, our estimate of \(C_o\) is reasonably close to the values proposed by Rowley (\(C_o=2.96\) km\(^2\) year\(^{-1}\)) and Parsons (\(C_o=2.51\) km\(^2\) year\(^{-1}\)), with the latter being based on an evaluation of consumption rates at subduction zones.

We follow Rowley [4] in concluding that discrepancies between Sclater et al. [3] and other studies probably result from the neglected back-arc spreading and marginal basins which may contribute ~0.38 km\(^3\) year\(^{-1}\) to the accretion rate estimates [27]. Adding this to our evaluation leads to \(C_o=3.230\) km\(^2\) year\(^{-1}\), which is almost identical to previous estimates [3,4,12,27].

### 3.2. Spreading rates

In order to evaluate the variations in spreading rate through geological time, we normalized the measured \(\{dA/dt\}\) quantities by the ridge lengths \(\{L\}\) at the time of production (Table 1). Not surprisingly, because most of the remaining ridge traces result from the breakup of Pangea, the ridges tend to lengthen with time, except for the Pacific ridges which are currently subducting. The spreading rates for each of the six oceanic areas considered, computed as \(\{dA/dt/L\}\), in mm year\(^{-1}\), are listed in Tables A and B and in the online appendix and are illustrated in Fig. 3. Because Pacific, Nazca, Cocos and Antarctic data are half-spreading rates, the spreading rates of the Atlantic and Indian oceans were divided by two in Fig. 3 (Table A in the online appendix).

Here again, a great variability between oceans is obvious (Fig. 3). During the Cretaceous (between chron M0 and 31; 120.4 to 67.7 Ma), the Atlantic ocean is characterized by an half-spreading rates of ~22 mm year\(^{-1}\) followed by a decrease to the present-day half-rate of ~14 mm year\(^{-1}\). In contrast, the Indian ocean rates first decrease from ~22 mm year\(^{-1}\) at chron M16 (139.6 Ma) to ~10 mm year\(^{-1}\) at M0 (120.4 Ma), and then remain stable at 18–20 mm year\(^{-1}\) except for the high rates of ~30 mm year\(^{-1}\) during the 67.7–47.9 Ma period (chrons 31 to 21), which corresponds to the time of rapid northward drift of the India plate. The Antarctic plate shows a different trend, with a slight decrease from ~20 to ~10 mm year\(^{-1}\) between M21 (147.7 Ma) and M0 (120.4 Ma), followed by a regular increase to present value of ~30 mm year\(^{-1}\). On average, the Pacific rates are twice as large as rates from the other oceans. Apart from the first two points (M29 and M25), which may be questionable in terms of accuracy, the Pacific rates tend to increase from ca. ~40 to ~75 mm year\(^{-1}\) from the end of the Jurassic to the beginning of the Cretaceous, between M21 (147.7 Ma) and M0 (120.4 Ma). Production rates decrease to ~30 mm year\(^{-1}\) at chron 25 (55.9 Ma) followed by a
significant, albeit scattered, increase to present-day values of 60–70 mm year\(^{-1}\). Such high values in recent times are corroborated by the Nazca and Cocos data.

From the above analysis, global spreading rates are dominated by the Pacific ocean. We however attempt to define a mean global spreading rate by averaging the values of Table A. This global mean (Table 2) was obtained by first averaging the Pacific, Nazca, and Cocos rates, in order to have an average spreading rate for the Pacific system. This value was then averaged with the Antarctic, half Indian and half Atlantic rates. The temporal evolution of spreading rates (Table 2, Fig. 3), computed only from currently observable elements, shows no significant variation since at least 155 Ma. There are slight variations in the mean half-spreading rates (Table 2). The mean before 83.5 Ma (chron 34) is: 25.0±1.3 mm year\(^{-1}\), 23.3±1.9 mm year\(^{-1}\) between 40.1 and 83.5 Ma (chrons 34 to 18), and 30.1±2.6 mm year\(^{-1}\) since 40.1 Ma. However, even if we consider these differences as significant within the 2\(\sigma\) error bars, the most striking and surprising feature is that the average spreading rates do not show a significant increase during the Cretaceous, as generally believed, but remain fairly stable at 25.9±3.3 mm year\(^{-1}\).

### 3.3. Seafloor generation rates

Beyond knowing the spreading rates themselves, the question of whether seafloor production underwent significant variations in surface is relevant to which length of ridges these rates applied to. Following Kominz [14], we assume that the total ridge length remained roughly constant with time (e.g. Tables 1a and b of Kominz [14]), and multiplying our constant spreading rate by a constant ridge length leads to a constant seafloor generation rate. However, we note that the higher spreading rates are biased by the larger production centers in the past, in particular, the Pacific system of ridges. This bias makes it necessary to re-evaluate the time-dependent production rate in another way, which implies, as discussed in Kominz [14], a number of hypotheses and assumptions. The main problem is to recover disappeared (subducted) ridges. From late Jurassic to present, two main oceanic ridges have undergone significant changes: the Tethys ridge(s), and the paleo-Pacific ridges.

The geologic evolution of the Tethyan domain was synthesized by Ricou [23], who also calculated the rotation parameters between the main plates. From these data, paleogeographic maps were drawn to reconstruct plate position through time, which included hypothesized ridges and subduction zones. We digitized these ridges from the maps [23], in order to obtain the Tethys ridge length over time (Table 3). For the paleo-Pacific ocean, we followed Engebretson et al. [22], and digitized the ridges separating the Izañagi (Iz), Farallon (Fa), Kula (Ku) and Pacific (Pa) plates, for the relevant time (Table 3). Fig. 4a illustrates the temporal evolution of total ridge length, following the above assumptions. Our results suggest global ridge lengths were slightly longer during the Mesozoic, which could also be the result of a poor determination of ridges lost by subduction. Despite

<table>
<thead>
<tr>
<th>Chron</th>
<th>(t) (Ma)</th>
<th>(D_t) (Ma)</th>
<th>Mean (\text{yr}^{-1})</th>
<th>(\sigma) (\text{yr}^{-1})</th>
<th>(n)</th>
<th>(r)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>0.0</td>
<td>9.7</td>
<td>33.2</td>
<td>20.6</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>9.7</td>
<td>10.4</td>
<td>27.1</td>
<td>15.7</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>20.1</td>
<td>13.0</td>
<td>30.8</td>
<td>21.1</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>13</td>
<td>33.1</td>
<td>7.0</td>
<td>29.2</td>
<td>21.8</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>18</td>
<td>40.1</td>
<td>7.8</td>
<td>21.4</td>
<td>11.6</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>21</td>
<td>47.9</td>
<td>8.0</td>
<td>25.9</td>
<td>11.3</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>25</td>
<td>55.9</td>
<td>11.8</td>
<td>22.9</td>
<td>10.0</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>31</td>
<td>67.7</td>
<td>15.8</td>
<td>22.8</td>
<td>7.4</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>34</td>
<td>83.5</td>
<td>36.9</td>
<td>24.5</td>
<td>15.4</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>M0</td>
<td>120.4</td>
<td>6.3</td>
<td>26.5</td>
<td>32.2</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>M4</td>
<td>126.7</td>
<td>5.2</td>
<td>22.9</td>
<td>13.6</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>M10</td>
<td>131.9</td>
<td>7.7</td>
<td>25.9</td>
<td>21.4</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>M16</td>
<td>139.6</td>
<td>8.1</td>
<td>24.9</td>
<td>13.2</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>M21</td>
<td>147.7</td>
<td>6.6</td>
<td>25.3</td>
<td>12.9</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>M25</td>
<td>154.3</td>
<td>6.2</td>
<td>45.2(a)</td>
<td>45.2</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>M29</td>
<td>160.5</td>
<td>19.5</td>
<td>30.7(a)</td>
<td>1 (--)</td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>&gt;M29</td>
<td>180.0</td>
<td>(--)</td>
<td>(--)</td>
<td>(--)</td>
<td>(--)</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Averages</th>
<th>(t) (Ma)</th>
<th>(D_t) (Ma)</th>
<th>Mean (\text{yr}^{-1})</th>
<th>(\sigma) (\text{yr}^{-1})</th>
<th>(n)</th>
<th>(r)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0–40.1 Ma</td>
<td>30.1</td>
<td>4</td>
<td>2.6</td>
<td>1.9</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>&gt;40.1 Ma</td>
<td>25.0</td>
<td>6</td>
<td>1.3</td>
<td>(--)</td>
<td>(--)</td>
<td></td>
</tr>
</tbody>
</table>

\(n\): number of entries in the statistics, \(\sigma\): standard deviation of the means.

\(a\) Excluded from the final means.
this, our evaluation is in quite good agreement with Kominz [14].

The next step in evaluating the surficial global seafloor production rates is to assign spreading rates to these ridges. It seems appropriate to assign the Pacific spreading rates, as we determined above, to the subducted paleo-Pacific ridges, especially because the older rates were already determined on parts of the segments that are almost entirely subducted. In contrast, the spreading rates of the Tethys sea are far less constrained, as this ocean has entirely disappeared. We thus decided to model seafloor productions in two ways: (1) the more conservative, and less favoured option applied the average global spreading rates we have determined above to the consumed ridges of the Tethysian and paleo-Pacific oceans; (2) the more realistic, and preferred option assigns the Pacific spreading rates to the paleo-Pacific ridges, and the Indian ocean rates to the Tethys ridges. The production rates $\frac{dA}{dt}$ of the subducted ridges are then added to the measured $\frac{dA}{dt}$ of the present-day oceanic crust.

The seafloor half-production rates from the present measurements, and from Tables 3 and 4 are illustrated in Fig. 4b. The total seafloor generation rates do not show large variations with time, in either option 1 or 2, and generally vary within $\pm 20\%$ about the mean. We note again that the average half-production rate of option 2 in the last 180 Ma, established at $1.298 \pm 0.284 \text{ km}^2 \text{ year}^{-1}$ (full rate=2.596 km$^2$ year$^{-1}$), compares well with the current-day production rate of 2.850 km$^2$ year$^{-1}$ evaluated above.

Table 3
Hypothesized subducted Ridges (km); interpolated values are in italics

<table>
<thead>
<tr>
<th>Chron</th>
<th>$t$ (Ma)</th>
<th>$Dt$ (Ma)</th>
<th>Tethys$^a$</th>
<th>Fa/Iz$^b$</th>
<th>Iz/Pa$^b$</th>
<th>Fa/Pa$^b$</th>
<th>Fa/Ku$^b$</th>
<th>Total Pacific ridges</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>0.0</td>
<td>9.7</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>5</td>
<td>9.7</td>
<td>10.4</td>
<td>780.3</td>
<td>780.3±78.0</td>
<td></td>
<td>780.3±78.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>20.1</td>
<td>13.0</td>
<td>967.9</td>
<td>967.9±96.8</td>
<td></td>
<td>967.9±96.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td>13</td>
<td>33.1</td>
<td>3.9</td>
<td>555.0</td>
<td>555.0±55.5</td>
<td></td>
<td>555.0±55.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td>18</td>
<td>40.1</td>
<td>4.9</td>
<td>444.0</td>
<td>444.0±66.6</td>
<td></td>
<td>444.0±66.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td>21</td>
<td>47.9</td>
<td>8.0</td>
<td>777.0</td>
<td>777.0±116.6</td>
<td></td>
<td>777.0±116.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td>25</td>
<td>55.9</td>
<td>9.1</td>
<td>999.0</td>
<td>999.0±149.9</td>
<td></td>
<td>999.0±149.9</td>
<td></td>
<td></td>
</tr>
<tr>
<td>31</td>
<td>67.7</td>
<td>12.3</td>
<td>3495.4</td>
<td>3495.4±524.3</td>
<td></td>
<td>3495.4±524.3</td>
<td></td>
<td></td>
</tr>
<tr>
<td>34</td>
<td>83.5</td>
<td>11.5</td>
<td>689.5±613.1</td>
<td>689.5±613.1</td>
<td></td>
<td>689.5±613.1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>95.0</td>
<td>15.0</td>
<td>8067.5</td>
<td>4910.6</td>
<td>4910.6</td>
<td>4910.6</td>
<td></td>
<td>9350.6±1153.7</td>
<td></td>
</tr>
<tr>
<td>110.0</td>
<td>10.4</td>
<td>12599.6</td>
<td>1852.6</td>
<td>1852.6</td>
<td>1852.6</td>
<td></td>
<td>17423.7±1551.0</td>
<td></td>
</tr>
<tr>
<td>M0</td>
<td>120.4</td>
<td>6.3</td>
<td>9428.3</td>
<td>9428.3</td>
<td>9428.3</td>
<td></td>
<td>17914.3±1541.4</td>
<td></td>
</tr>
<tr>
<td>M4</td>
<td>126.7</td>
<td>5.2</td>
<td>8925.5</td>
<td>8925.5</td>
<td>8925.5</td>
<td></td>
<td>18221.8±1543.2</td>
<td></td>
</tr>
<tr>
<td>M10</td>
<td>131.9</td>
<td>3.1</td>
<td>8510.4</td>
<td>8510.4</td>
<td>8510.4</td>
<td></td>
<td>18474.8±1548.2</td>
<td></td>
</tr>
<tr>
<td>135.0</td>
<td>4.6</td>
<td>8264.0</td>
<td>4013.8</td>
<td>4013.8</td>
<td>4013.8</td>
<td></td>
<td>18628.0±1553.0</td>
<td></td>
</tr>
<tr>
<td>M16</td>
<td>139.6</td>
<td>5.4</td>
<td>8531.5</td>
<td>8531.5</td>
<td>8531.5</td>
<td></td>
<td>18852.2±1562.1</td>
<td></td>
</tr>
<tr>
<td>M21</td>
<td>145.0</td>
<td>2.7</td>
<td>8845.6</td>
<td>8845.6</td>
<td>8845.6</td>
<td></td>
<td>18093.0±1829.2</td>
<td></td>
</tr>
<tr>
<td>M25</td>
<td>154.3</td>
<td>6.2</td>
<td>8693.5</td>
<td>8693.5</td>
<td>8693.5</td>
<td></td>
<td>18093.0±1829.2</td>
<td></td>
</tr>
<tr>
<td>M29</td>
<td>160.5</td>
<td>19.5</td>
<td>8585.9</td>
<td>8585.9</td>
<td>8585.9</td>
<td></td>
<td>18093.0±1829.2</td>
<td></td>
</tr>
<tr>
<td>M29</td>
<td>180.0</td>
<td></td>
<td>9479.4</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>


Uncertainties on the total are evaluated by assuming a 20% uncertainty on Tethys estimates, and a 15% uncertainty for the other disappeared ridges.

$^a$ Following Ricou [23].

$^b$ Following Engebretson et al. [22].

$^c$ Subducted Nazca isochrons after Royer et al. [1].
However, it seems that option 2 coincides with an increase in production rate, which reaches $42 \pm 20\%$ of the mean rate, between M4 (126.7 Ma) and M0 (120.4 Ma), at the beginning of the Cretaceous. Because the period between M0 and M4 is short, a slight shift in time of one or both of these chron could lead to a drastic change in the generation rate. Furthermore, the rates then rapidly decrease during the Cretaceous, falling down to minimum values (~1 km$^2$ year$^{-1}$) around chron 31 (67.7 Ma) and chron 18 (40.1 Ma). A similar evolution is obvious using option 1, yet with lower amplitudes, in particular during the Cretaceous. Another quite surprising feature, obvious from both options 1 and 2, is that the production rates have tended to increase since 40.1 Ma.

3.4. Oceanic crust production rates

It is straightforward to evaluate magmatic fluxes at oceanic ridges provided surface production rates and crustal thicknesses are known. The chemical composition (major and trace elements) of oceanic basalt is important because it can be quantitatively linked to mantle temperature [28,29,31]. Humler et al. [5] investigated ancient crustal compositions from the Atlantic, Pacific and Indian oceans. They found that the chemical composition of oceanic crust older than 80 Ma was statistically different from the modern one. In order to explain this observation, they proposed that the mantle temperature was 50 $^\circ$C hotter prior to 80 Ma, whereas the mean mantle temperature of the mantle after 80 Ma is identical to the temperature beneath the modern oceanic system [32]. On the other hand, the amount of melt in the mantle beneath an oceanic ridge is controlled by the amount of pressure release that each individual parcel of mantle has experienced (e.g. [28,29]). The higher the mantle temperature, the more melt produced, and hence the greater the thickness of the oceanic crust (e.g. [30]). Regardless of the origin of mantle temperature variations [32,33], the chemical data suggest that crustal thicknesses were 1–2 km thicker before 80 Ma (~9 km) than at present (~7 km) [5]. Klein and Langmuir [28] found a positive global correlation between the extent of melting, derived from the major-element chemistry of basalts, and seismically determined crustal thicknesses.
In effect, using the average seismic crustal thickness of oceanic crust from White et al. [35] and Nagihara et al. [36], one obtains an average of 6.6 ± 0.8 km ($n=24$) for 0–15 Ma oceanic crust, and of 7.4 ± 1.0 km ($n=34$) for crust older than 80 Ma [5]. Albeit ~1 km smaller, these estimates are in good agreement with those deduced from the chemical composition of oceanic basalts. Moreover, the geochemical results suggest that the crust was not grown thicker [37,38] but rather that old crust was thicker at the time of its formation.

As a consequence, despite the constant surface generation rates reported here at oceanic ridges for the last 180 Ma, production rates changed due to crustal thickness variations linked to mantle temperature variations. Thus, for 0–80 Ma, we propose that the magmatic flux at oceanic ridges was 18.1 ± 3.9 km$^3$ year$^{-1}$, being the product of the average surface generation rate 2.596 ± 0.568 km$^2$ year$^{-1}$ multiplied by a 7-km-thick crust. In contrast, during the Mesozoic, the magmatic flux was higher by about 30% ($23.3 ± 5.1$ km$^3$ year$^{-1}$ for a 9-km-thick crust). Accounting for seismically determined thicknesses, this difference could be reduced to ~15%.

Note that our analysis has neglected the volumes of oceanic plateaus. An evaluation of oceanic plateaus production was given by Larson [19], following the work of Schubert and Sandwell [34]. The mean oceanic plateau production rate was about 3 km$^3$ year$^{-1}$ since 150 Ma [19], some 60% lower than the excess production rate at ocean spreading ridges during the Mesozoic. We therefore conclude that, although oceanic plateau production is an important factor, the effect of mantle temperature variations on ridge production is far from being negligible, and may be the major cause in the balance of mass production within ocean basins.

### 4. Discussion and conclusion

We have calculated/measured the area of oceanic crust per unit age ($dA/dt$) as a function of age, from the isochron database of Royer et al. [1] and Müller et al. [2]. Our results are in fairly good agreement with the previous estimates of Sclater et al. [3], even though we used a different and probably more precise method of computing surface areas on the Earth, as well as an updated data set of Isochron positions on the Earth, and magnetic timescales.

We then computed the spreading rates within each oceanic basin by dividing these surfaces by the ridge lengths as measured from the isochron lengths. These calculations show that the global average half-spreading rates have remained constant since at least 150 Ma, at 25.9 ± 3.3 mm year$^{-1}$. Making some hypotheses on subducted paleo-Pacific ridges (following [22]) and Tethys ridges (following [23]), and on the velocity on these ridges, we finally propose a roughly constant seafloor surface generation rate since 180 Ma, within ±20% variation around a mean (half) value at 1.298 ± 0.284 km$^2$ year$^{-1}$ (Table 4). There may exist a slightly higher production rate at the beginning of the Cretaceous, but this observation relies on hypothetical ridges, with unknown velocities, and, as a consequence, should be considered questionable.
Our evaluation of the seafloor surface generation half rate is notably lower than the full production of 3.4 $\text{km}^2\text{ year}^{-1}$ proposed by Rowley [4], because we did not take into account small ocean basins. However, in agreement with Rowley [4], but based on a different method, we conclude that spreading rates remained constant since at least 180 Ma. Importantly, the fact that both studies reach similar conclusions using quite different methods clearly shows that the initial assumptions used in the integration of Rowley [4] are valid: (1) the destruction of oceanic floor at subduction zones is uniformly distributed with age, and (2) from present to 180 Ma, the oldest age of oceanic crust remained constant at 180 Ma.

If significant, the evolution of seafloor surface generation rates during the last 180 Ma is in significant disagreement with the results of Kominz [14], who suggested spreading rates in the Cretaceous, between 120 and 80 Ma, attained two to four times the present-day value, followed by a regular decrease since then. Apart from the fact that we found an increase in production rates since 40 My, which is not assumed by Kominz [14] model, the Cretaceous appears to us as a period characterized by a particularly quiet, and fairly slow, seafloor generation rate. We point out, however, that the determinations of Kominz [14] are primarily based on Pacific plate reconstruction models, involving a significant amount of subducted oceanic crust. The Kominz model [14] also assumes symmetric spreading and neglects the effects of ridge jumps during the Cretaceous LNS. It is also bothersome that the inferred high spreading rates take place during the LNS, a period for which, by definition, no precise description of seafloor spreading history can be obtained. Another limiting factor of this work is that the timescales were poorly defined at that time. We therefore conclude that our evaluation, which is based on much more conservative methods and assumptions, and probably more accurate timescales, provides a more accurate description of seafloor generation rates.

From our results, no clear correlation appears between spreading or surface generation rates and the 1st order variations in sea-level [17]. Following Heller et al. [20] and Hardebeck and Anderson [21], who proposed Pangea breakup as an alternative mechanism, we also conclude that the 1st order sea-level changes are not primarily controlled by the average seafloor spreading rates, as proposed by Hays and Pitman [13]. However, instead of Pangea breakup, we suggest that variations in oceanic crust thickness related to mantle temperature variations since the Mesozoic could account for the sea-level changes.

Finally, our analysis shows that despite a constant spreading history during the past 180 Ma, the crust production at ridges could actually be higher during the Mesozoic because the mantle was hotter. Although this mechanism differs from previous suggestions, it provides a consistent alternative hypothesis for the high CO$_2$ concentration in the Earth’s atmosphere during the Cretaceous.

Acknowledgements

We thank L.E. Ricou, J. Dyment and V. Courtillot for their thoughtful insights and discussions all along this work. J.Y. Royer and J. Francheteau are thanked for their helpful discussions. W. Crawford, S. Gilder and J. Ludden have greatly improved the final version of this manuscript. We also thank D. Müller, D. Rowley and an anonymous reviewer for their comments and corrections.

Appendix A. Supplementary data


References


