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STRESS-INDUCED CHANGES IN PLATE DENSITY, VAIL SEQUENCES, EPEIROGENY, AND SHORT-LIVED GLOBAL SEA LEVEL FLUCTUATIONS

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Abstract. The stratigraphic record contains clear evidence that broad regions have experienced rapid changes in sea level unrelated to changes in glacial ice volume (e.g., third- and higher-order Vail sea level variations, continental foundering). We propose that many of these changes can be caused by stress-induced changes in plate density. Stress changes produce significant changes in the density of the crust and lithosphere (a point missed in previous investigations) and propagate across even the largest plates in less than 30,000 years. Lithospheric plates interacting at existing boundaries can produce stress-related density changes sufficient to cause several meters change in plate elevation; these may account for many of the regressions and transgressions seen in the stratigraphic record. The creation of new rifts could increase plate compression enough to cause ~50 m of plate subsidence. Plate elevation changes of up to -200 m could result from increased plate compression during continental collisions. A particularly enigmatic kind of rapid nonglacial global (NGG) sea level change is a coupled ~50 m sea level fall (regression) and roughly equal rise (transgression) occurring in less than 10^6 years. These couplets, associated with black shales and marine extinctions, can be explained by the elastic snapback attending the rapid formation of a new rift. Isostatic disequilibria along the new rift depresses the seafloor sufficiently to cause a ~50 m fall in sea level. Mantle flow restores isostatic equilibrium along the rift axis and erases the fall in ~60,000 years. Unusually intense hydrothermal circulation along the new rift during the snapback promotes anoxic bottom water conditions and deposition of black shales. A rapid drop in global sea levels reduces the area of ocean bottom within the photic zone, causing overpopulation, food exhaustion, and extinctions. The connections between changes in plate stress and density have many implications. Perhaps the most far reaching is that tectonic events on the 3/4 of the globe that is covered by oceans, as well as the 1/4 that is subaerial, are recorded in sedimentary strata.

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INTRODUCTION

Global sea level changes have fascinated geologists since they were first proposed by Suess [1906] to account for apparent worldwide correlations in the stratigraphic record. Interest has been heightened by the development and wide application of seismic stratigraphic methods [Vail et al., 1977; Haq et al., 1987] and the observation that certain types of rapid, nonglacial, sealevel changes correlate with marine extinctions better than any other geologic factor [Hallam, 1989].

Sealevel changes can be divided into five types, as illustrated in Figure 1. There are gradual changes in sea level of up to a few hundred meters caused by changes the volume of mid-ocean ridges [Pitman, 1978; Heller and Angevine, 1985]. These shifts in sea level occur over hundreds of millions of years. There are rapid changes in sea level caused by the growth and dissipation of continental glaciers. These can be up to ~100 m and occur over ~10,000 years (melting) to 100,000 years (accumulation) [Pitman, 1978; Donovan and Jones, 1979]. Between these understood extremes lie a full spectrum of sea level rates and magnitudes caused by vertical motion of the continents and global nonglacial changes in sea level related in some fashion to tectonic processes. Some of the changes are regional and rapid [e.g., Officer and Drake, 1985]. It has been particularly difficult to find suitable explanations for third- and higher-order Vail cycle sealevel changes (<5 m.y. sea level cycle) and the rapid epeirogenic and nonglacial global (NGG) sea level changes shown specifically in Figure 1 [Schlanger, 1986].

As far as we know there have been three quantitative suggestions to explain the Vail sea level changes and regional epeirogenic movements and one quantitative suggestion for rapid truly global changes. Cloetingh et al. [1985] showed that regional stress changes of 10^{16} dyn/cm (e.g., a 4 kbar stress change acting on a 25 km thick portion of the lithosphere, or a 2.5 kbar stress change acting on a 40 km thick section, etc.) could cause 50-80 m elevation changes at the margins of a basin. The greater response occurred in thin lithosphere (new oceanic crust). Officer and Drake [1985] suggested chaotic mantle convection might be responsible for the regional epeirogenic and perhaps NGG sealevel changes. Sabadini et al. [1990] showed that true polar wander could cause sealevel changes of opposite polarity in the northern and southern hemispheres that would be rapid enough to accommodate third-order Vail cycles. These mechanisms are all possible, but for different reasons none are fully satisfactory.

The mechanism proposed by Cloetingh et al. [1985] applies within a band about 100 km wide surrounding the flexural depression caused by a sediment load. Stress-induced elevation changes vary from zero at the inner and outer edges of the band to a maximum in the center of this band. Thus highly variable sea level changes would be expected

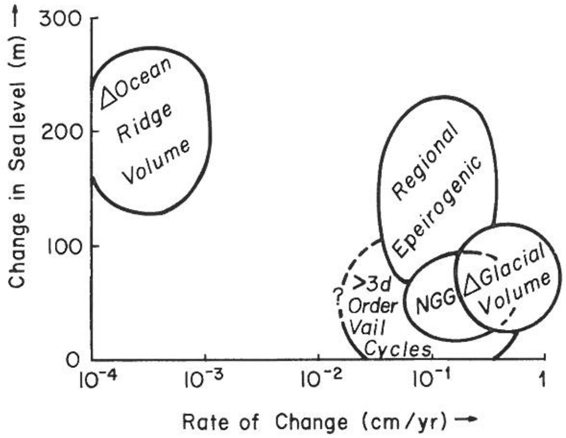


Fig. 1. Magnitude and rate of sea level changes inferred from the stratigraphic record. Third- and higher-order Vail cycles, rapid epeirogenic sea level changes, and nonglacial global (NGG) changes in sea level have been particularly difficult to understand.

along a coastline that cuts across the load margin. Their mechanism cannot explain the broad, uniform continental foundering and emergence sometimes observed in the stratigraphic record. Regarding elevation changes tied to mantle convection, even if chaotic changes in mantle convection occur, it is far from clear that they will occur rapidly enough to explain the rapid nonglacial sea level changes. Although the hemispheric asymmetry makes the Sabadini et al. [1990] hypothesis easily testable, it is far from clear that episodes of polar wander are frequent enough or of the right character to account for third-order Vail cycles, and polar wander cannot account for epeirogenic movements of a single continent.

We show below that stress changes in the lithosphere similar in magnitude to those appealed to by Cloetingh et al. [1985] change the lithosphere density enough to cause frequent changes in isostatic plate elevation of a few meters and occasionally cause elevation changes of ~50 m when new rifts develop. In previous treatments this point has been entirely missed. It has been uniformly assumed that stress-induced density changes would be far too small to have any appreciable isostatic effect. Surprisingly, this can be shown not to be the case, especially if the magnitude of plate stress changes are similar to the 4 kbar changes in a 25 km thick lithosphere appealed to by Cloetingh et al., who have made a good case for stress changes of this magnitude. In the case of stresses produced by continental collision, changes in plate elevation up to 200 m could be produced. The density of the entire plate is uniformly affected, and thus the founderings are truly regional and uniform in nature. Stress changes thus have the potential not only to account for the frequent rapid changes in plate elevation suggested by third- and higher-order Vail sedimentation cycles but also for episodes of widespread and rapid continental submergence and emergence.

If the stress and strain changes are very rapid, as might be allowed by the sudden rifting of a plate, the elastic snapback of the suddenly ruptured plate will produce transient regression-transgression couples of a magnitude and time scale suitable to explain the NGG variations in Figure 1. Plate stress changes thus could explain the third- and higher-order Vail sediment cycles and the epeirogenic and NGG sea level changes in Figure 1. Sudden rifting is admittedly speculative, but it is supported by some evidence and is worth considering for this reason and because it is very hard to think of any other viable mechanisms for producing global sea level cycles on the NGG time scale.

In what follows we first summarize stratigraphic evidence for sea level change. We then show how Vail cycles and epeirogenic and NGG sea level changes could be caused by stress changes in lithospheric plates and the snapback transients that would attend the rapid rifting of large plates.

SEA LEVEL CHANGES, EXTINCTIONS, AND RELATED PHENOMENA

The Phanerozoic record contains abundant evidence of frequent (every ~2 m.y.) episodic changes of sea level in the form of sedimentary cycles and onlap/offlap relationships. The sea level cycles are evidenced by stratigraphic sequences enclosed between unconformities that show clearly on reflection seismic records. There is little agreement on the magnitude of the sealevel change. The most widely publicized estimates have been those of the Exxon group in their well-known sealevel curves [Vail et al., 1977; Haq et al., 1987]. Long-term changes were calibrated against changing oceanic ridge volume since the mid Cretaceous, while short-term changes were based on the shape of their coastal onlap/offlap curve. Large amplitudes of 100 m or more were inferred for many of their so-called type 1 unconformities on the assumption that coastal onlap and sea level shift below the shelf edge. Christie-Blick et al. [1991] have strongly challenged this assumption and question the validity of the technique, while Hallam [1988] could not recognize any type 1 unconformities in the Jurassic of Europe that could convincingly be related to a eustatic fall. By far the biggest sea level fall recognized by the Exxon group was in the mid-Oligocene. Vail et al. [1977] estimated this to be nearly 400 m within the duration of a planktonic faunal zone, about a million years, but failed to explain why this estimate was heavily weighted in favor of North Sea data [Hallam, 1981, p. 134]. Haq et al. [1987] reduced this figure to 150 m. The best estimate of Miller et al. [1987] using a variety of techniques was 30-90 m. The evidence seems to indicate ~60 m sea level fall in <1 m.y.

Two other approaches applied to the deposits of ancient epicontinental seas are exemplified by Hallam [1978] and Holser and Magaritz [1987].

Epicontinental seas are likely to have been extremely shallow over extensive stretches, such that even a small change of sea level would effect drastic changes of environment of a type that cannot be readily compared with today's pericontinental seas and open oceans, with a correspondingly greater effect in the invertebrate faunas [Johnson, 1974; Hallam, 1981, chapter 5]. Hallam [1978] attempted to estimate the amount of sea level change for the European Jurassic by assigning plausible values to the likely depth of deposition of the varied facies types and concluded that even the deepest-water facies could have been deposited within 100 m of sea level, an inference supported more recently by the recognition by Wignall [1989] of evidence of storm activity in the organic-rich shales of the Kimmeridge Clay of Dorset, generally considered to be relatively deep-water facies. Consequently, sea level changes of only a few tens of meters would be sufficient to account for the facies changes observable up the stratigraphic sequence. A 50 m sea level rise or fall within half an ammonite zone, say, about 0.5 m.y. would be a reasonable estimate for the biggest events.

The above discussion assumes that the unconformities bounding stratagraphic sequences are produced by subaerial erosion. Changes in sea level are basically measured by changes in elevation along the surface of the unconformity. Arguments revolve around the paleoslope of the unconformity surface (hypsometry), how to decompact, and so on. Another, completely different interpretation of the unconformity surfaces has recently been proposed [Pitman and Golovchenko, 1991; W. C. Pitman, Large, rapid sea level changes, submitted to Earth and Planetary Science Letters, 1990]. Pitman proposes that the unconformities are produced by submarine (not subaerial) erosion. Erosion profiles are tied to shoreline position and are very sensitive to sealevel change. The swings in sea level required to produce the observed unconformity-bounded Vail stratigraphic sequences are reduced by an order of magnitude, from 50 to ~5 m.

Another debated aspect of unconformity-bounded sedimentary sequences is whether the causative sea level cycles are local (due to epeirogenic movements of the land masses relative to the sea) or global (due to global or eustatic changes in sea level). Although sediment sequences have been interpreted as having been caused by eustatic changes in sea level [Vail et al., 1977; Haq et al., 1987; Goodwin and Anderson, 1985], it is more likely that the majority are regional phenomena related to epeirogenic activity [e.g., Officer and Drake, 1985; Cloetingh et al., 1985].

Having said this, nevertheless, based on intra- and intercontinental biostratigraphic correlations of marine deepening and shallowing events and corresponding transgressions and regressions, some of the sea level changes appear to have been genuinely under eustatic control [Hallam, 1984]. Of particular interest are what have been termed regression-transgression couplets [House, 1989], where a comparatively rapid sea level fall has been quickly followed by a correspondingly rapid rise characteristically associated with the spread of anoxic bottom water. Such events often correlate with mass extinctions of marine invertebrates [Hallam, 1989], which provides an additional argument in favor of truly global events. If the events were restricted in extent to limited regions, organisms could have retreated to widespread refugia at times of increased environmental stress and no mass extinction would have ensued.

Holser and Magaritz [1987] used an indirect method to estimate the amplitude of the regression-transgression sea level cycle across the Permian-Triassic boundary. Using information from the literature on the distribution of marine and nonmarine strata in conjunction with a modern hypsometric curve, they arrived at a figure of about 280 m for the latest Permian fall and earliest Triassic rise. This figure seems excessive. A fall as drastic as this would have caused all the epicontinental seas to disappear and the Late Permian sediments to be exposed to significant erosion. Regional stratigraphic reports reveal, however, a considerable measure of agreement that physical indications of hiatuses in marine sequences across the Permian-Triassic boundary are virtually absent [Logan and Hills, 1973], and carbon isotope profiles through such sequences give no indication of abrupt changes that could imply stratigraphic gaps in the southern Alps [e.g., Holser et al., 1989], but elsewhere the profiles do show discontinuities [Baud et al., 1989]. The sea level changes are suggested by nondeposition (missing biostratigraphic zones) in some marine sections and by nonmarine deposition in others. There is no physical evidence of erosion, however. It is likely that the assumption of unchanging continental hypsometry through time is unjustified, and there are considerable uncertainties about how to interpret the simple distinction of marine and nonmarine facies, though this is probably reliable enough to indicate a striking qualitative change across the boundary (Figure 2a). The time span of fall and recovery, regardless of magnitude, was several million years. The latest Permian fall in sea level was less rapid than the earliest Triassic rise.

The Cretaceous-Tertiary (K-T) boundary has been widely accepted by generations of stratigraphers as one of the biggest events in the Phanerozoic record. A sea level fall in the latest Maastrichtian followed by an earliest Danian rise is indicated by data from Tunisia, Alabama, and Italy. Variations between studies and localities indicate the uncertainties in defining global sea level changes. Elevated percentages of sporomorphs, land-derived organic matter and an inshore shallow water dinoflagellate group, and calculated sedimentation rates using age-calibrated bioevent data on the most complete Cretaceous-Tertiary boundary section known, near El Kef in Tunisia [Keller, 1988], for example, indicate a peak regressive pulse lasting about 2×10^4 years exactly at the K/T boundary [Brinkhuis and Zachariasse, 1988]. The regressive pulse occurred at the end of a longer-term regressive trend and was

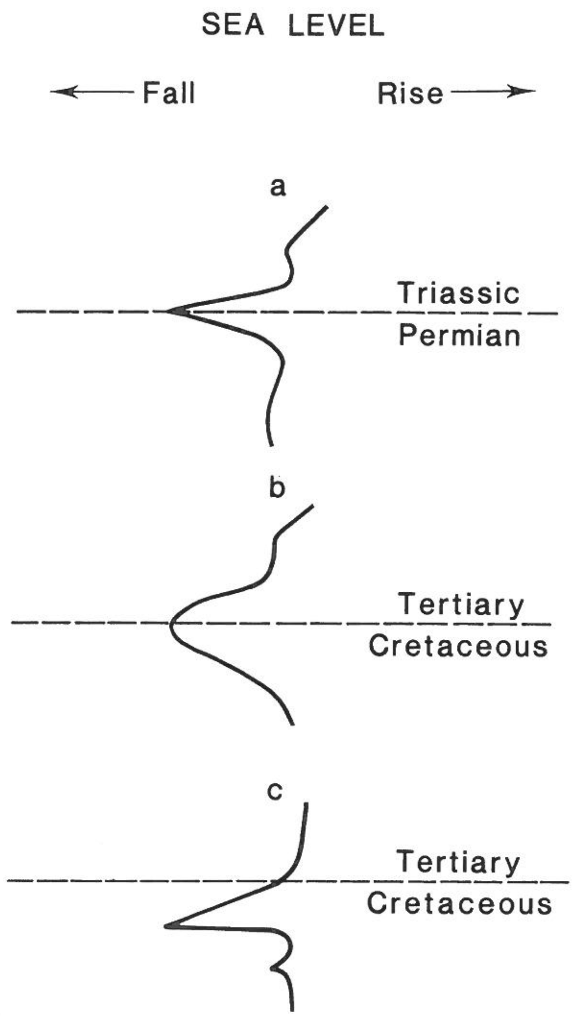


Fig. 2. Regression-transgression couplets are a particularly significant kind of nonglacial global sea level change that correlate with marine mass extinctions better than any other geologic factor. The sea level changes occur over less than a million years and have magnitudes of ~50 m. The three listed cases are from (a) Holser and Margaritz [1987, Figure 8], (b) Brinkhuis and Zachariasse [1988, Figure 7], and (c) Donovan et al. [1988, Figure 9] (Figure 9 of Donovan et al. [1988] is after Haq et al. [1987]).

immediately followed by a sea level rise in the early Danian. Sea levels did not reach Maastrichtian levels until a long time later (Figure 2b). The Exxon group studied an excellent K/T boundary section at Braggs, Alabama, and concluded that the main regressive event took place not at but shortly before the boundary and estimated the sea level fall and rise to be about 80 m [Haq et al., 1987]. Further study of the Braggs section has led, however, to the event being placed closer to the boundary and the sea level change estimate being revised upward to about 130 m, making it appear a much more significant event than in the earlier-published curve [Donovan et al., 1988] (Figure 2c). A longer lasting shallowing-deepening event at the K/T boundary can be inferred for the classic locality of Gubbio, Italy, from the clay mineral data of Johnsson and Reynolds [1986]. On the reasonable assumption that the marked increase of kaolinite in a 3 m interval across the boundary is best interpreted as due to a sea level regression [Hallam, 1987] a duration of ~2 x 10^5 years is suggested.

Clearly, there is considerable uncertainty about both the amount and duration of rapid nonglacial sea level changes of all types. There seems little doubt, however, that there have been a number of global sea level changes of ~50 m magnitude that occurred in <~1 m.y. at rates ≥0.01 cm/yr. For most of the Phanerozoic it is not plausible to invoke glacioeustasy to account for these short-term episodic sea level changes because of the absence of evidence of major polar ice caps. This is especially clear for the Devonian and Jurassic, but the strata of both systems contain abundant evidence for such events [House, 1985, 1989; Johnson et al., 1985; Hallam, 1978, 1988]. The ascent of mantle plumes could give rise to epeirogenic movements and associated surface volcanicity [Loper et al., 1988] but, whereas uplift of parts of the ocean floor could clearly cause marine transgressions of the continents by displacement of seawater, it is difficult to see how comparable upward displacement of continents could lead to regression on a global scale or to regression-transgression couplets. It is these couplets (Figure 2) that must form the key to interpretation. By some means there must be a mechanism to restore sea level to something close to its original position in a geologically brief period of time. The extent of global sea level change need never have been much more than 50 m, and many such events could have been appreciably less than this.

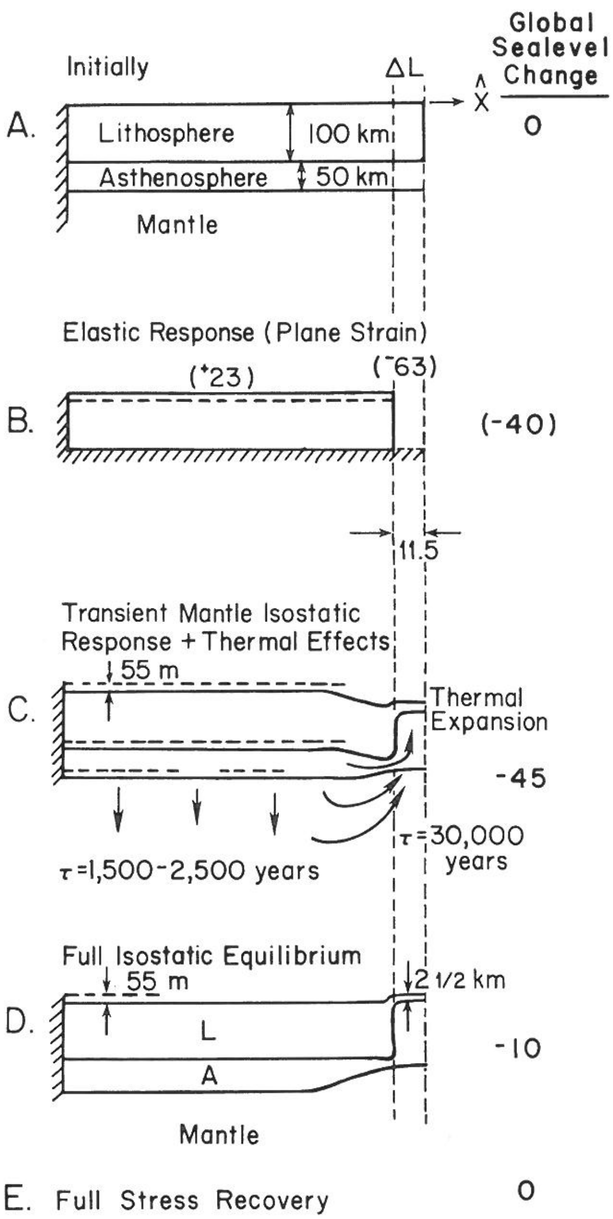
STRESS-INDUCED CHANGES IN LITHOSPHERE DENSITY AS A CAUSE OF THIRD- AND HIGHER-ORDER VAIL CYCLES, EPEIROGENY, AND EUSTATIC SEA LEVEL CHANGE

Changes in plate stress can account for third- and higher-order Vail sea level changes and epeirogeny; transient mantle flow can account for eustatic NGG changes in sea level. The basic concept is simple. Plate tectonic factors change the stress at the margins of a plate. The stress changes the density of the plate and changes the freeboard accordingly. The margin stresses take about 10,000-30,000 years to diffuse into the interior of the plate, and thus the time scale of strain and sea level change is appropriate to explain the higher-order Vail cycles and epirogenic and NGG sea level changes. Here we concern ourselves with the situation after all transients are completed and full isostatic equilibrium is restored and inquire after the longer-term magnitude of the sea level changes both on the plate on which stress has been changed and globally. The following section considers transient effects.

Because boundary conditions are observed to change mainly at the affected margin, it is most appropriate to assume plane strain. For plane strain, the strain parallel to a new rift (in the $\hat{y}$ direction of Figure 3) is not changed, deviatoric stress in the $\hat{x}$ direction perpendicular to the rift axis is reduced to near zero, and the vertical stress remains unchanged. Under these conditions the strain in the $\hat{x}$ and $\hat{z}$ directions can be deduced from equations given by Turcotte and Schubert [1982, p. 110 ff]:

$$\Delta\varepsilon_x = \frac{\delta L}{L} = \frac{(1+\nu)(1-\nu)}{E}\Delta\sigma_x\ , \tag{1a}$$

$$\Delta\varepsilon_z = \frac{\delta h_L}{h_L} = \frac{-\nu(1+\nu)}{E}\Delta\sigma_x\ . \tag{1b}$$



L is the length of the plate, ν is Poisson's ratio, E is Young's modulus, h_L is the thickness of the crust and lithosphere, and $\Delta\sigma_x$ is the change in horizontal stress in the horizontal x direction.

The change in volume per unit volume of the plate is the algebraic sum of the volume changes due to thickening of the plate plus those due to opening the rift gap divided by the initial volume, which equals minus the change in change in average density of the lithosphere divided by the average lithoshere density and also the sum of the two strain components given above:

$$\frac{\delta h_L L^2 + \delta L\, h_L L}{h_L L^2} = \frac{-\delta\rho_L}{\rho_L} = \frac{(1+\nu)(1-2\nu)}{E}\Delta\sigma_x, \tag{1c}$$

where ρ_L is the average density of the crust and lithosphere. Compressing ($\Delta\sigma_x$ negative) the plate thus thickens it (1b) and increases its density (1c). Had plane stress conditions been assumed (e.g., had it been assumed that stress parallel to the new rift rather than strain was unchanged), the expression for change in density would have been the same as above with the $(1 + \nu)$ factor removed. Plane strain assumptions thus produce a density change about 27% larger than plane stress assumptions. The plate response will lie between these two assumptions but closer to plane strain. The differences are not large. Plane strain is assumed in the calculations that follow.

The change in ocean depth, δd, over an isostatically equilibrated and contracted oceanic plate

Fig. 3. Cartoon illustrating the development of a rift opened by the sudden rupture and elastic snapback of an initially stretched plate. The plate is 10,000 km wide, and the horizontal stress change is 4 kbar over 25 km. (a) Condition of the stretched plate before rupture. (b) Response if the elastic plate rests on a frictionless, perfectly rigid substrate. The 23 km gap would cause a 63 m fall in sea level; the elastic thickening of the plate would cause a 23 m rise. (c) Isostatic adjustment under the compressed and densified plate will be rapid (1500-2500 year time constant) and lead to a 55 m subsidence (rather than a 23 m rise) relative to adjacent plates on which stress has not changed. Asthenospheric material will redistribute locally into the rift. Isostatic adjustment of the mantle along the rift will occur with a time constant of ~30,000 years. The result will be a transient depression of sea level of ~-45 m. (d) After ~60,000 years full isostatic equilibrium will be established. The final result will be a 55 m depression of the stress-changed plate (relative to adjacent plates) and a -10 m global change in sea level. (e) If the plate is restressed to its initial condition, the 55 m subsidence will be recovered and the 10 m fall in global sea level erased.

relative to unaffected neighbor plates is easily derived (as shown in Appendix A) by requiring equal mass in prisms of equal cross section to some depth of compensation in the asthenosphere before and after elastic relaxation (the definition of isostasy). The result is

$$\delta d = \frac{\rho_L - \rho_a}{\rho_a - \rho_w}\,\delta h_L + \frac{h_L}{\rho_a - \rho_w}\,\delta\rho_L \,, \tag{2}$$

where ρ_L is the density of the lithosphere, ρ_a the density of the asthenosphere, and ρ_w the density of water. Substitution of (1b) and (1c) yields

$$\delta d = -\left(\frac{\rho_L - \rho_a}{\rho_a - \rho_w}\,\nu(1+\nu) + \frac{\rho_L}{\rho_a - \rho_w}\,(1+\nu)(1-2\nu)\right)\frac{h_L\,\Delta\sigma_x}{E}\,.$$

For $\nu = 0.269$ [Bullen, 1965], an average density of the oceanic lithosphere $\rho_{Lo} = 3.23$, an average density of the continental lithosphere, $\rho_{Lc} = 3.05$, a density of the asthenosphere, $\rho_a = 3.18$, [e.g., Watts and Steckler, 1981; Royden et al., 1980], and $\rho_w = 1.0$, the first term in the above equation is less than 2.4% of the second (<1% for oceanic lithosphere). Changes in lithosphere thickness do not affect plate elevation significantly and can be neglected in comparison to lithosphere density changes.

Assuming the lithosphere consists of strong and weak parts and that plate stresses elastically deform the strong interval(s) with thickness h_{Ls}, the equation for δd given above can be rewritten

$$\delta d = -\frac{\rho_L}{\rho_a - \rho_w}\,(1+\nu)(1-2\nu)\,\frac{h_{Ls}\,\Delta\sigma_{xs}}{E}. \tag{3}$$

In (3) it is important to appreciate that h_{Ls} is the thickness of the strong part of the lithosphere at any particular instant and $\Delta\sigma_{xs}$ is the stress that acts over that interval. The product $h_{Ls}\Delta\sigma_{xs}$ is the change in force acting on the plate which is assumed constant after any change. The thickness of the strong part of the lithosphere, h_{Ls}, may change with time if the lithosphere is viscoelastic, but the product $h_{Ls}\Delta\sigma_{xs}$ and δd will not change.

Taking $E = 1.6 \times 10^{12}$ dyn/cm^2 and the other parameter values as given above, (3) reduces to

$$\delta d = -\begin{pmatrix} 5.5 \\ 5.0 \\ 3.0 \end{pmatrix} \times 10^{-13}\ \mathrm{cm^2/dyn}\ h_{Ls}\Delta\sigma_x \quad \begin{matrix} \mathrm{OP} \\ \mathrm{C@SL} \\ \mathrm{C{>}SL} \end{matrix}\,, \tag{4}$$

where the first line is for oceanic plate (labeled OP), the second line for continents at sea level, and the third for continents above sea level. If we take the same force change $h_{Ls}\Delta\sigma_{xs} = -10^{16}$ dyn/cm used by Cloetingh et al., [1985], (4) shows that compression (negative $\Delta\sigma_{xs}$) will cause a 55 m subsidence of an oceanic plate relative to adjacent similar plates on which stress has not changed and a 50 m subsidence of a continental plate initially at sea level. These relative elevation changes are surprisingly large. They are similar in amplitude to those Cloetingh et al. showed would be produced by changes in regional stress interacting with the flexural depression associated with a sediment load. Plate force changes during continental collision could be -4 x 10^{16} dyn/cm, as discussed in Appendix B. Plate compression caused by continental collisions therefore could cause up to 200 m of regional subsidence relative to plates on which stress has not changed.

The subsidence (relative to adjacent plates) caused by stress change and the sea level change over the plate on which stress has changed is independent of the size of the plate. The size of the plate will, however, control how much the regional change in relative plate elevation affects global sea level. Stress changes on a large plate can cause significant changes in global sea level.

Assuming the plate affected by changes in stress is and remains below sea level, the general change in sea level, ΔO, as seen from continental land masses not involved in the stress changes (see Appendix A) is

$$\Delta O = -(1 - \rho_w/\rho_a)\,L^2\delta d/A_{ocean}, \tag{5}$$

where δd is given by (3) or (4). Taking $A_{ocean} = 3.62 \times 10^8$ km^2 and a large plate with area 10^8 km^2, (5) shows that subsidence (rise in sea level) on a large plate newly compressed by a force of 10^{16} dyn/cm will produce a general fall in sea level on all continental land masses not affected by the stress change about 19% as large as the increased ocean depth over the plate itself. A 55 m increase in water depth over the compressed plate will thus correspond to a general fall in sea level of about 10 m. For the maximum subsidence of ~200 m that could be caused by continental collision, the sea level change could be ~-38 m.

These calculations show that plausible stress changes in a lithospheric plate alter its density enough to cause significant changes in the elevation of that plate with respect to neighbor plates on which stress has not changed. If large ocean-covered plates are affected, significant changes in general sea level also result. The relative elevation changes of the plate on which stress has changed are comparable in magnitude to those which would be caused by the interaction of regional stress and flexure at the margins of basins or other loads [Cloetingh et al., 1985] but differ from the flexural elevation changes in that they apply to large portions of or entire plates rather than to just the relatively narrow fringe areas around flexural depressions.

The magnitude of the relative change in plate elevation depends on the magnitude of the force change. If a new rift decoupled a plate from its former subduction zone, the increased compression seen by the plate could easily be 10^{16} dyn/cm (see Appendix B), and the resulting plate subsidence whould be ~55 m. The changes in regional elevation associated with new continental collisions could be considerably larger. New rifts and new continental collisions are rare events, and these large changes in relative plate elevation will be infrequent. Changes in relative plate elevations can also be produced by stress changes at existing plate boundaries, however. Here stress changes will be of a few hundred bars (rather than a few kilobars), and thus changes of plate elevation of ~5 m may be expected. Intuitively, it seems reasonable that these "boundary jostling" changes in plate stress, density, and elevation could be frequent enough to account for the sea level regressions and transgressions seen in the sedimentary record and known as third-order (and higher) Vail cycles.

Plate jostling may produce significant changes in global sea level depending on how the stress changes coordinate. Change in global sea level could be similar to the changes in the elevation of individual plates, for example, if a significant percentage of the world's oceanic plates were affected by stress changes of the same sign. In this case, or in the case of large stress changes on a single large oceanic plate, some interesting counterintuitive sea level relations would result. A significant rise in sea level on the stress changed plate or plates would coincide with a fall in sea level on all other plates. The pattern of sea level change could provide a powerful test of the hypothesis that stress changes cause sea level changes.

THE RATE OF STRAIN RECOVERY

How long does it take to alter the stress in a plate? Consider stress release by rifting. From (1a) and the parameter values used previously, a drop in tensile stress of 4 kbar on a large plate will produce a horizontal strain of $\frac{\delta L}{L} = 2.3 \times 10^{-3}$. Thus rifting of a 10,000 km wide plate strained to near failure will open a 23 km wide rift if broken in the middle. The minimum time required to recover the strain across an initially stressed plate after a sudden rupture is controlled by how rapidly the lithosphere can move across the top of the viscous asthenosphere.

The rate of contraction of an instantly rifted plate can be calculated in a fashion analogous to that used to calculate strain release after a major earthquake [e.g., Turcotte and Schubert, 1982, p. 374]. A force balance equation equates the elastic stress (with plane strain boundary conditions) to the stress required to shear the asthenosphere, assuming no flow conditions at the bottom of the asthenosphere (so-called Coulette flow conditions). This leads to a diffusion equation of elastic displacement:

$$\frac{\partial u}{\partial t} = \frac{h_a h_{Ls} E}{(1-\nu^2)\,\eta} \frac{\partial^2 u}{\partial x^2}\,. \qquad (6)$$

In (6), u is the elastic displacement of the plate measured from the fixed end at the left side of Figure 3, h_a is the thickness of the asthenosphere, h_{Ls}, as before, is the thickness of the strong part of the lithosphere that is stressed and controls the strain and strain recovery, η is the viscosity of the asthenosphere, E is Young's modulus, and ν is Poisson's ratio. Note that we assume in (6) that portions of the lithosphere outside the rigid interval characterized by h_{Ls} will be entrained and move sympathetically with the rigid parts with no added resistance.

Immediately after rifting, recoverable displacement increases linearly from zero at the left margin of Figure 3a to 11.5 km at the rift. The boundary conditions are no displacement, u = 0, at the left margin and no stress along the rift. The problem is analogous to cooling a rod with a triangular heat distribution (11.5°C initially at the center, corresponding to the rift, and 0°C at the left side) subject to boundary conditions of 0°C at the left margin and insulating conditions at the rift. The solution is given by Carslaw and Jaeger [1976, p. 98]. Carslaw and Jaeger's solution shows that if the width of the rifted plate is L, 90% of the strain is recovered in a time τ:

$$\tau = \frac{(1-\nu^2)\,\eta L^2}{h_a h_{Ls} E}\,. \qquad (7)$$

Over 35% of the strain will be recovered in 0.1 τ, and 70% of the strain will be recovered in 0.4 τ. For L = 5000 km, h_{Ls} = 25 km, h_a = 50 km, $\eta = 10^{20}$ poise, E = 1.6×10^{12} dyn/cm^2, and ν = 0.262 [e.g., Bullen, 1965, p. 233; Cathles, 1975], τ = 30,000 years.

Most of the parameters in the decay time calculation are uncertain by a factor of 2. The only point we wish to make is that the pullback of the plate is likely to occur over a few tens of thousands of years. The pullback of a plate from a newly formed rift thus occurs on the time scale needed to explain the rapid nonglacial sea level regressions discussed above. Since no plate is much larger than 10,000 km square, the response time of the system of lithospheric plates will be similar to the 30,000 years estimated for the single large plate discussed above.

Rifting need not be rapid. It could proceed by propagator migration, for example, at a rate similar to the rebuilding of stress after rupture. Judging from the Exxon curves, the time to restress a rifted plate (or other equivalently sized plate or plates) to near failure and return sea level to its levels prior to rifting appears to be a few million years [Haq et al., 1987]. If the rifting takes place over a million year time scale (or even over a 100,000 year time scale), isostatic equilibrium can be maintained throughout the process, and the isostatic analysis in the previous section will apply.

GLOBAL CHANGES IN SEA LEVEL ASSOCIATED WITH ISOSTATIC DISEQUILIBRIA AT NEW RIFTS

It is interesting, however, to consider the consequences if the rifting were abrupt. If rifting takes place over a time scale less of than 100,000 years, isostatic disequilibrium near the new rift can contribute significantly to the changes in sea level. These changes can be larger and more abrupt than the global sea level changes associated with the stress-induced changes in plate elevation and are superimposed upon them. They could therefore be especially significant for the survival of marine species.

The magnitude of possible disequilibrium changes in sea depth can be estimated by considering the compressed plate to rest on a perfectly rigid substrate as illustrated in Figure 3. Note that compression is caused both by the elimination of tension (as when a new rift cuts a trench from its former plate) and by the smaller effects of ridge push. The magnitude of global or eustatic sea level change is proportional to the length of the rift, the width of the rifted plate, and the magnitude of the stress (not force) change. We consider here an admittedly large 10,000 by 10,000 km plate. Rifting of a 10,000 km wide plate strained to near failure will open a 23 km wide rift during the elastic snapback. If this rift is 10,000 km long, a volume of lithosphere will be transported from the rift area sufficient to change ocean depth everywhere by -63 m. This assumes the lithosphere at the new rift is 100 km thick. The rift gap will of course be largely filled by asthenospheric material drawn from adjacent areas. The effects on sea level will be unchanged, however, because depression of the adjacent areas from which asthenosphere is drawn will exactly compensate seafloor elevation and seawater displacement in the rift gap. At the same time, for a rigid substrate there would be a 23 m rise in sea level caused by the elastic thickening of the retracted plate.

Asthenospheric material will also be entrained and moved out of the rift area by the movement of the overlying lithosphere. Since Coulette flow should be approached a few hundred kilometers from the rift and redistributions of asthenosphere near the rift do not in themselves affect mean ocean depth, the amount of material removed from the rift area can be estimated from the volume of a wedge 23 km wide at the top (equal to the lithosphere displacement), 0 km wide at 50 km greater depth (the thickness of the asthenosphere), and 10,000 km long. This volume of asthenosphere material is sufficient to produce a -32 m change in ocean cover over all other plates. The increased depth of the ocean near the rift by elastic retraction of the asthenosphere and asthenosphere entrainment is therefore sufficient to decrease the depth of the oceans worldwide by about 96 m. Note we have ignored the increased thickness of the plate caused by elastic retraction and have ignored the fact that material removed from the rift area by asthenosphere entrainment will thicken the adjacent asthenosphere. This is justified because, as discussed below, isostatic adjustment will be very rapid under the plate as a whole. As a result, the thickened asthenosphere under the plate will not change sea level, and the thickened lithosphere will actually contribute (not detract) from the sea level fall as discussed in the stress change section above.

The isostatic response of the mantle is illustrated in Figure 3c. In general terms the densified lithospheric plate and thickened asthenosphere will cause isostatic subsidence of the top of the mantle under the plate as a whole, while the mantle under the gap and neighboring thinned asthenosphere will rise. Because the dimensions of the compressed plate are large, the isostatic response of the mantle will be rapid. Isostatic equilibrium under the plate as a whole will be approached with an exponential decay time intermediate between the 2500 year response time observed for the Wisconsin glacial load that covered a 3300 km diameter area of North America and the 1500 year response time inferred for the ocean basins to the postglacial meltwater influx by the lack of a Holocene high sea level on continents relative to ocean islands (see, for example, Cathles [1975]).

The mantle response to unloading at the new rift will be much slower because response time is inversely proportional to the size of the load. The rift zone effectively (e.g., allowing for some local asthenosphere redistribution as in Figure 3c) unloaded by lithosphere movement and asthenosphere thinning is at most about 100 km wide, which is about an order of magnitude less than the 1100 km glacial load in Fennoscandia, whose melting caused isostatic uplift with a time constant of 4400 years [Cathles, 1975]. Taking into account the different geometry of the loads (equidimensional for Fennoscandia versus a highly elongate strip) decreases the response time by a factor of 0.7 [Cathles, 1975, Appendix VI], so that the response time of the mantle to the gap unloading should be about 33,000 years.

Comparison of the above plate and gap response times to the 12,000-30,000 year response time for 70-90% strain recovery in the lithosphere after rifting indicates that the plate itself should remain in isostatic equilibrium during strain recovery but that a strong degree of isostatic disequilibrium should be produced in and near the rift.

If rift depression of 96 m is ~50% erased by mantle response as the gap is opened, the transient change in sea depth would be ~-50 m. Isostatic equilibrium between plates means that mantle material displaced from the gap will cause a globally uniform uplift in regions not directly involved and no change in continental sea level but the sea depth changes will cause isostatic adjustments between oceans and continents that reduce their impact on continentally observed sea level changes by 70% (e.g., as indicated in (5); see Appendix A for discussion). We therefore expect rapid rifting of a 10,000 x 10,000 km oceanic plate to contribute an additional 35 m to the sea level fall produced by strain recovery in the rifted plate. Sea level will thus fall a total of about 45 m on a ~30,000 time scale and return to -10 m after an additional 30,000 years for any ocean-covered plate, as shown in Figure 4a. Full recovery to prerifting sea levels will occur in a few million years as the rifted plate (or an equivalent plate or plates) is restressed. If the same size continental plate is rifted suddenly but remains above sea level, we would expect a global ~35 m fall in sea level in ~30,000 years followed by full recovery to prerift levels after an additional 30,000 years.

Figure 4b combines the subsidence due to stress-induced density changes and those due to transients along the rift and shows the change in sea level relative to continental land masses on a 10,000 x 10,000 km plate as a function of time since rifting. If the plate is largely oceanic or continental with the continents at sea level before rifting, there will initially be a slight (5-10 m) fall in sea level. This is because the 50-55 m subsidence of the plate will slightly exceed the 45 m transient drop in sea level caused by the dynamic, nonisostatic depression along the rift. As the rift depression disappears, the plate will experience a 40-45 m transgression. This transgression will be removed only if the plate is restretched to its prerift state. If the plate is a large continental mass with elevation well above sea level, there will be almost no initial sea level change because the 35 m transient drop in sea level will be almost exactly canceled by the 30 m elastic foundering of the plate. After the rift transients have been erased by mantle flow, the plate will experience a ~30 m transgression which will be removed only if and when the plate is restretched.

It should be noted that the changes in sea level produced by the new ridge are comparatively small. A 23 km wide ocean ridge 2 1/2 km above the seafloor will displace enough water to increase ocean depth globally by 1.6 m and lead to a 1.1 m rise in sea level with respect to the continents. This is a negligible change in sea level compared to those caused by the lithosphere retraction and asthenosphere entrainment that will be associated with the same event, and for this reason we have ignored the ridge effects.

DISCUSSION AND CONCLUSIONS

The numbers given above should of course be taken as illustrative. The calculations we present are simplified and designed to demonstrate the hypothesis that stress changes could cause significant regional and global sealevel changes. There is considerable uncertainty in many of the parameters, not the least because known geologic phenomena can change the values. For example, a plume could change the thickness and viscosity of the asthenosphere and, with time, the thickness and yield strength of the lithosphere. The greatest uncertainty is the magnitude of the stress changes that can occur in the lithosphere. The yield strength of mature lithosphere and the thickness of the lithosphere when it yields are not well known, and it is unclear whether it is plausible to suppose that the lithosphere could fail rapidly over ~10,000 km. We have made estimates, approximations, and even some hypotheses that can certainly be validly debated.

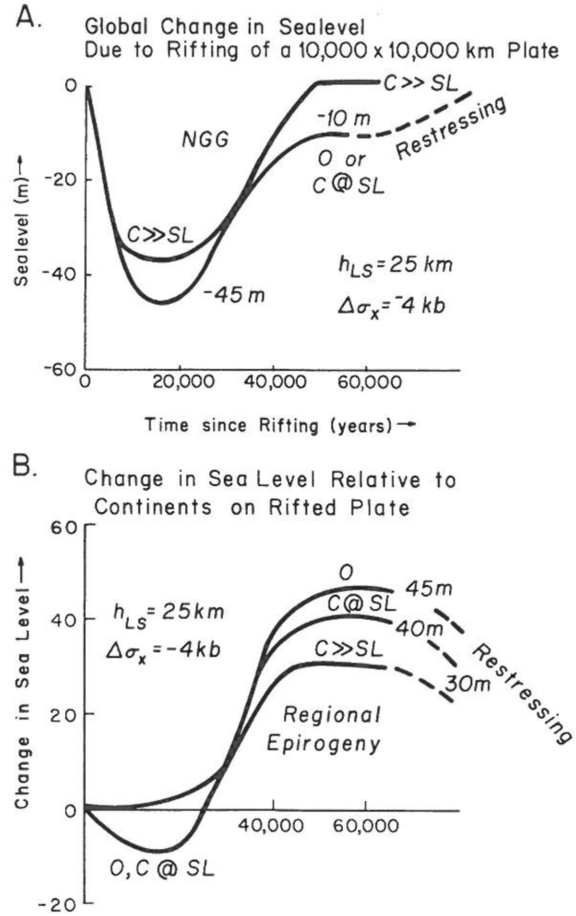


Fig. 4. (a) Changes in sea level associated with the rifting of a 10,000 km x 10,000 km plate. The -45 m transient sea level change is appropriate in magnitude and time scale to explain the regression-transgression couplet (NGG) sea level changes in Figures 1 and 2. O indicates the plate is an oceanic plate, C@SL indicates the entire plate is a continent initially at sea level which then subsides below sea level, and C>>SL indicates a continent initially sufficiently above sea level that it is never ocean covered. (b) Change in sea level relative to continents is given as a function of time for the same event in Figure 4a. Plate subsidence has been combined with changes in global sea level to give the change in sea level relative to continents. The symbols are the same as in Figure 4a. Note that at least one small continent must be on the oceanic plate to record the sea level changes plotted. See Appendix A for a discussion of how sea level curves would differ on oceanic islands.

The first and most basic point we wish to make is that stress changes produce density changes which, through isostasy, cause significant changes in plate freeboard. Stress-induced density changes have previously been completely overlooked. Stress changes of -200 bars (compression) acting over a 25 km interval of lithosphere thickness can cause broad regions to uniformly subside ~3 m. The amount of subsidence depends only on the force change (stress change-thickness product) and is independent of the size of the plate. A few hundred bars stress change might be produced by the normal jostling of plates along established boundaries. To the extent that shoreline regressions and transgressions in the stratigraphic record can be accounted for by a few meters of sealevel change, the continual jostling of plates can account for the character of the stratigraphic record in a very natural way.

Larger changes in plate freeboard can infrequently be produced by the new rifting of mature plates or new continental collisions. The increase in the ocean depth over plates compressed by such events (relative to unaffected plates) could be ~55 and ~200 m, respectively. If the compressed plates were large and ocean covered, significant changes in global (eustatic) sea level would result. If the plate affected were 10,000 x 10,000 km in size, for example, the change in global sea level would be ~-10 and ~-38 m, respectively. New rifting events and continental collisions are known to have occurred in the geologic past, and it is possible that stress changes associated with these events produced the continental-scale sea level inundations that have traditionally been attributed to epeirogenic movements of the continents. It is interesting in this regard to note that Bally [1982, p. 334] has emphasized the continent-wide character of subcrop unconformities and has noted that they "appear to be preferably formed at the inception of a major collisional phase...". We are in the process of investigating, for example, the relation of the Zechstein transgressions in Europe to tectonic events at the Permian-Triassic boundary. Further discussion is outside the intended scope for this paper.

If Vail sequences in general require ~100 m changes in sea level rather than a few meters, the stress change hypothesis must be considered only a long shot possibility. It is possible that large stress changes associated with rifts, collisions, or other kinds of tectonic events (major jostlings) could be frequent enough to account for third- and higher-order Vail sequences, but it is not likely.

The most speculative part of this paper is the suggestion that extensive rifts can on occasion develop rapidly enough to produce significant isostatic disequilibria along the rift during their "elastic snapback". Sudden failure lacks an obvious mechanism but is suggested by some geologic evidence and could account for some otherwise enigmatic observations. Our analysis of the consequences of rapid rifting shows that transient sea level changes associated with the sudden rifting of a large (10,000 x 10,000 km) plate can produce a global sea level regression of ~45 m followed, on a ~30,000 year time scale, by a nearly equal transgression. This regression-transgression couplet is of exactly the right magnitude and duration to explain the origin of the type of nonglacial global (NGG) sea level changes associated with mass marine extinctions and black shales.

Small-scale examples of aspects of the processes we appeal to can be identified. Sleep and Rosendahl [1979] proposed that a dynamic loss of fluid head in asthenospheric material upwelling to fill the gap left by plate separation at ocean ridges accounted for 1 1/2 km deep axial grabens found at slow spreading ocean ridges. This mechanism can explain the ~3 1/2 km subsidence along the Green Tuff Belt in Japan that preceded simultaneous formation of the Kuroko massive sulfide deposits along this ~1000 km long rift in the middle Miocene [Cathles et al., 1983]. The dynamic isostatic disequilibrium is demonstrated in Japan by the fact that when the Green Tuff rift aborted, the depressed areas returned to near prerift elevations. The Green Tuff rift thus responded dynamically as envisioned in the snapback model of sea level change.

The drop in global sea level associated with the rifting in Japan is small (few meters at most). The rift involved a small back arc plate. Eustatic sea-level variations related to regression/transgression cycles are noted at about the right time (13.8 ma [Haq et al., 1987]); these variations could be related to events in Japan. The massive sulfide deposits found in the Green Tuff Belt of Japan are about an order of magnitude larger than those found so far on the ocean floor. The greater size may be due to the formation of a larger magma chamber during rifting of an island arc or due to the increased efficiency of sulfide deposition allowed by the anoxic bottom water conditions that were produced by particularly vigorous hydrothermal discharge in the narrow sea with restricted circulation produced by the rifting event.

Schlische [1990] recently discussed similar phenomena associated with the opening of the Atlantic. Extrusion of a distinctive high-titanium, high-magnesium quartz-normative tholeiitic flood basalt occurred simultaneously in all the northern basins of the Newark Supergroup (a failed Atlantic rift), from the Culpeper Basin in Maryland and Virginia to the Bay of Fundy Basin in Nova Scotia. Ar^{40}/Ar^{39} dating indicates the outpourings occurred over an interval of less than 670,000 years at the Triassic-Jurassic boundary, 201 m.y. before present. About 1 km of total extension can be estimated from the heave and dip of normal (listric) faults and from the thickness of the flood basalt unit. The Great Dike of Rhodesia is another possible example of a snapback event. It extends for over 500 km, is several kilometers wide, and formed, as far as can be determined, with two satellite dikes in one event 2650 ± 16 m.y. before present [Cahen et al., 1984].

Other phenomena may suggest snapback. A sudden rift failure with tens of kilometers of elastic snapback is the kind of event that could produce a funnel-shaped magma chamber of the type described by Pallister and Hopson [1981] in Oman. Pore water convection is required to produce the narrow magma chambers observed at steadily spreading ridges along the East Pacific Rise [e.g., Cathles, 1990]. Similar pore water convection would keep a very rapidly growing magma chamber narrow at its base (near the Moho) where the pore fluids impinge on the chamber. Boundary layer èffects would reduce heat transport at higher levels, promoting the the funnel shape. The result could be a funnel-shaped magma chamber with span approximately equal to the snapback gap. As another example, if the crust is compressed by the elastic contraction of the strong parts of the lithosphere, fluids might be expelled from entire regions during rifting or collision events, not just from the areas deformed by overthrusting ("squeegee" zone of Oliver [1986]).

Phenomena related to rapid plate rifting have interesting implications for the black shales and marine extinctions associated with nonglacial global regression-transgression couplets. First, the pullback associated with sudden rifting of a stretched plate will be largely achieved in 12,000 years. The rate of seafloor creation along a 10,000 km rift would be ~20 km^2/yr, or ~7 times the current global rate of seafloor generation at all present ridges. The resulting increased hydrothermal circulation through the new oceanic or thinned continental crust could introduce reduced hydrothermal waters into the ocean (or shallow narrow sea in the case of a rifted continent) rapidly enough to help drive the bottom waters anoxic and account for the association of sea level regression-transgression couplets and black shales. Second, rapid rifting of an oceanic plate could, under certain circumstances, have a profound affect on marine life. The most vulnerable condition would be one in which the continental land masses had extensive shallow seas. New rifting of a large oceanic plate under these circumstances could drop sea level and dramatically reduce the area of shallow parts of the shelf (<50 m deep) that lie within the photic zone. Photic bottom dwellers would suddenly have to survive in a drastically smaller living space, causing species extinctions [see Wyatt, 1987].

Several predictions of the stress change-density-sea level hypothesis could provide a means of testing it. The prediction that rifting can cause sealevel transgression on one plate and regression on all others seems particularly testable. The prediction of the model that a plate that has recently undergone rapid rifting and rapid strain recovery should have an unusually thin asthenosphere at the rift also might be testable. This is not the relation that one would normally expect. Rapid forced extension associated with a strain recovery pullback might also exaggerate thermal perturbations along the rift, since viscosity depends strongly on temperature. These might be recorded as a pattern of basins and unconformities along the margins of continental rifts.

A particularly far-reaching implication of the hypothesized connection between sea level and changes in plate stress is that events on oceanic as well as continental plates may be recorded in the sedimentary record. Sedimentary strata provide the most complete and continuous record of geologic events. If the jostling of plates is recorded in Vail sea level fluctuations, stratigraphic sequences may provide insights into other geologic events in plate interiors such as dikes, faulting, vertical movements, and so on. If new rifting and collision events on the 3/4 of the globe that is ocean covered (as well as the 1/4 that is not) can be extracted from the sedimentary record using the often nonintuitive relations sketched above, we would have a nearly complete record of the major tectonic events that have affected the Earth. Such a record could show temporal patterns, for example, that would provide new insights into mantle dynamics.

APPENDIX A: ISOSTATIC AND MASS BALANCE CONSIDERATIONS

Mantle and ocean mass must be conserved. After any disequilibria produced by dynamic processes has decayed, isostatic equilibrium must exist everywhere and, since this means mantle pressure is unperturbed, mantle and ocean volume must be conserved. Isostatic equilibrium and conservation of mantle and ocean volume are most easily enforced by requiring isostatic equilibrium between adjacent lithospheric plates and using conservation of ocean and mantle mass as auxiliary conditions.

Isostatic equilibrium between a destressed plate and an identical adjacent plate that is not destressed requires mass in the columns shown in Figure A1 be equal to some depth of compensation in the asthenosphere. From Figure A1, isostasy can be expressed

$$\rho_W d + \rho_L h_L = \rho_W(d - \delta h_L - \varepsilon) + (\rho_L + \delta\rho_L)(h_L + \delta h_L) + \rho_a \varepsilon, \qquad \text{(A1)}$$

where ε is the change in elevation of the lithosphere asthenosphere boundary as shown in Figure A1.

Fig. A1. Illustration of isostasy requirements between destressed plate (right) and an identical plate not de-stressed (left); h_L is the thickness of the lithosphere, d is the depth of the ocean, ρ is the density of water, the lithosphere, or the asthenosphere (subscripts w, L and a, respectively), and δ indicates a small change in these parameters.

The first term on the left of (A1) cancels the first term on the right. The second term on the left is canceled by a term on the right. Neglecting second-order terms (e.g., products of δh_L and $\delta\rho_L$), yields

$$\varepsilon = -\frac{\rho_L-\rho_w}{\rho_a-\rho_w}\delta h_L - \frac{h_L}{\rho_a-\rho_w}\delta\rho_L . \tag{A2}$$

The change in ocean depth, δd , equals $-(\delta h_L + \varepsilon)$, so

$$\delta d = \frac{\rho_L-\rho_a}{\rho_a-\rho_w}\delta h_L + \frac{h_L}{\rho_a-\rho_w}\delta\rho_L . \tag{A3}$$

This corresponds to equation (2) in the text. Note that at this point we have not considered the change in sea level over both plates caused by changes in elevation of the stress-changed plate. This would lead to identical isostatic adjustment and change in sea depth over both. The term δd in (A3) thus should be considered the change in water depth over the plate on which stress has been changed relative to an adjacent plate whose stress has not changed and which remains below sea level.

Isostatic equilibrium must also be maintained between continental and oceanic plates after any global change in sea level. If the depth of the oceans has changed globally by an amount δd_G, isostatic balance between the ocean floor and continents requires that the ocean floor move relative to the continents by an amount δd_I such that $\rho_w\delta d_G + \rho_a\delta d_I = 0$. Conservation of ocean mass relates δd, the increase in ocean depth over a plate of dimension L on which stress has changed, and δd_G:

$$\delta d_G = \frac{-L^2\delta d}{A_{ocean}} . \tag{A4}$$

The change in sea level relative to the continents, ΔO, is given by the sum of δd_G and δd_I:

$$\Delta O = \delta d_G + \delta d_I = \left(1 - \frac{\rho_w}{\rho_a}\right)\delta d_G \tag{A5}$$
$$= -\left(1 - \frac{\rho_w}{\rho_a}\right)\frac{L^2\,\delta d}{A_{ocean}} .$$

Note that mantle mass is conserved by a small uniform change in mantle radius, which, because it affects continent and ocean alike, does not affect sea depth or sea level relative to the continents.

Changes in sea level will depend on the point of observation. For example, using the parameter values given in the text, for δd = 55m on a 10,000 x 10,000 km plate, δd_G = -55 (1/3.62) = -15 m, δd_I = 15 (1/3.18) = 5 m, and ΔO = -10 m. Thus ocean depth on the stress-changed plate will increase 55 m relative to adjacent ocean plates on which stress has not changed. The actual ocean depth increase is 40 m because ocean depth has decreased globally 15 m. A dipstick island on the stress-changed plate will record a 40 m increase in sea level. A dipstick island on an adjacent oceanic plate whose stress has not changed will record a 15 m fall in sea level. Continental land masses will record a 10 m fall in sea level. The continental record of sea level change differs by 5 m from the oceanic islands because of the isostatic adjustment of the ocean floor relative to the continents. In the text, sea level changes are generally considered to be measured relative to continental coastlines.

APPENDIX B: STRESS CHANGES IN LITHOSPHERIC PLATES

The thickness of the stressed layer, h_{Ls}, can be assessed in a number of ways. First, it can be inferred from the observed flexural rigidity of the oceanic or continental lithosphere taking the elastic parameters of the crust determined seismically. This generally yields values of effective lithosphere thickness of about 40 km.

Alternatively, the thickness of the strong part of the lithosphere can be estimated from Brierlie's law and a creep law for olivine. This approach also provides an estimate of the differential stress at which the lithosphere will yield under tension or compression. Heat flow is a critical parameter in this method. Dewey [1988] has recently compiled a number of these curves for heat flows of 1 cal/(cm^2 s). It can be deduced from his figures that the lithosphere is strongest in compression or tension over an ~25 km thick interval (h_{Ls}) and that the maximum sustainable tensile and compressive lithospheric forces, $h_{Ls}\Delta\sigma_{xs}$, are 10^{16} dyn/cm and -3 x 10^{16} dyn/cm, respectively. Thus the maximum stress change any plate can undergo is about 4 x 10^{16} dyn/cm, or 16 kbar applied to a 25 km thick portion of the lithosphere.

Finally, a variety of observations suggest that stress in the lithosphere can reach levels of 2 to 4 kbar (see reviews by Cloetingh et al., [1985] and Cloetingh [1988a, b]). These stress estimates are consistent with those that could be produced by ridge push and trench pull stresses (each ~0.5 kbar acting over a 100 km thick lithosphere), provided these forces are focused on a 25 km thick portion of the lithosphere [e.g., Dewey, 1988]. For the parameters in the text, which were selected for their compatibility with models that can successfully account for the subsidence histories of extensional basins, unresisted slab pull can be large. The unresisted pull per unit length of trench equals the product of the density contrast between the average lithosphere and asthenosphere times the thickness of the lithosphere (100 km) times the length of the slab (700 km) which equals ~1.8 x 10^{16} dyn/cm for a density contrast of 0.05 g/cm^3 at the trench reduced linearly to zero at 700 km depth. This is equivalent to 7 kbar acting on a 25 km thick strong interval of the lithosphere. Trench pull, although mostly neutralized by resistive forces along the subduction zone, can clearly be large.

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