A new multilayered model for intraplate stress-induced differential subsidence of faulted lithosphere, applied to rifted basins

R. T. van Balen, Y. Y. Podladchikov, and S. A. P. L. Cloetingh
Tectonics/Structural Geology Group, Faculteit der Aardwetenschappen, Vrije Universiteit, Amsterdam

Abstract. In-plane horizontal stresses acting on predeformed lithosphere induce differential flexural vertical motions. A high-precision record of these motions can be found in the sedimentary record of rifted basins. Originally, it was proposed that rifted basins experience flank uplift and basin center subsidence in response to a compressive change of in-plane stress, which agrees well with observed differential motions. Subsequently, published models predicted that the vertical motions may be opposite because of the flexural state of the lithosphere induced by necking during extension. However, the total, flexural and permanent, geometry of the lithosphere underlying the rifted basin is the controlling parameter for the in-plane stress-caused vertical motions. The largest part of this preexisting geometry is caused by faulting in the uppermost brittle part of the crust and ductile deformation in the underlying parts of the lithosphere. We present a new multilayered model for stress-induced differential subsidence, taking into account the tectonically induced preexisting geometry of the lithosphere, including faults in the upper crust. As continental lithosphere may exhibit flexural decoupling due to a weak lower crustal layer, the new multilayer in-plane stress model discriminates the geometries of the separate competent layers. At a basin-wide scale, the model predicts that a compressive change of in-plane force results in basin center subsidence and flank uplift, confirming the original hypothesis. Compared to all previous models, the new model requires a lower horizontal stress level change to explain observed differential vertical motions.

1. Introduction

The global stress field, inferred from, for example, earthquake focal mechanism solutions, in situ stress measurements, and borehole breakout analyses, has persistent trends, which are uniform over large areas and can be correlated to plate motions [Cloetingh and Wortel, 1985; Zoback et al., 1989; Zoback, 1992; Müller et al., 1992; Richardson, 1992; Coblenz and Richardson, 1995; Lindholm et al., 1995]. The intraplate stress field originates largely from the integrated result of lithospheric density heterogeneities and processes operating at plate boundaries, that is, slab pull and ridge push [Wortel et al., 1991; Richardson, 1992; Coblenz and Richardson, 1995]. Paleocorestress analyses (from stylolites, fault strataions, etc.) demonstrate that intraplate stress orientations and magnitudes can change on a timescale of a few million years [Philip, 1987; Letouzey, 1986; Bergerat, 1987; De Ruig et al., 1991; Cosso et al., 1991]. Furthermore, folding of oceanic [Cloetingh and Wortel, 1985; Stein et al., 1989; Beckman et al., 1996] and continental lithosphere [Stephenson and Cloetingh, 1991; Martinod and Davy, 1992; Iffort and Agarwal, 1996] points to very large magnitudes of intraplate stresses of up to several hundred megapascals [Stephenson and Cloetingh, 1991; Cloetingh and Burov, 1996].

In rifted sedimentary basins, a changing horizontal intraplate stress level causes flexural subsidence and uplift, as long as the stress level is lower than the limit defined by strength of the lithosphere and its weakness zones (see section 6). The induced relative sea level changes are recorded as onlap and offlap events in the basin stratigraphy [Cloetingh et al., 1985]. Therefore flexural uplift and subsidence caused by intraplate stress level changes provide a tectonic mechanism for sea level changes which interferes with truly eustatic sea level changes. As a consequence, flexural deformations due to continuously changing intraplate stress magnitudes and directions can provide a plausible explanation for relative sea level changes which apparently occurred during the Mesozoic [Cloetingh et al., 1989], as inferred from (seismo) stratigraphic data [see Haq et al., 1987]. Thus far, no other known mechanism (e.g., glacioeustasy) can satisfactorily explain the inferred high-frequency Mesozoic sea level changes [Pitman and Golovchenko, 1983], although there is some evidence for the existence of polar ice [Ziegler, 1990].

Numerical forward modeling of the North Sea basin [Kooi and Cloetingh, 1989], the Barents Sea basin [Reemst et al., 1994], the Beaufort-Mackenzie basin [Tang and Lerche, 1992] and the Pannonian basin [Van Balen et al., 1998] has demonstrated that a changing level of intraplate stress can explain the observed differential subsidence or relative sea level changes. These rifted basins all experienced enhanced subsidence in the basin's center in response to compression (or a relative sea level rise) during their postrift phase, confirming the predictions of Cloetingh et al. [1985].

In this paper a new model for stress-induced subsidence and uplift in rifted basins during their postrift evolution is presented, based on recent insights into the rheology and flexural behavior of the lithosphere (e.g., Burov and Diamant,...
1995] and the influence of existing faults on its flexural state [Kusznir et al., 1991; Spudich and Preobrazhenskii, 1996]. In contrast to previous work on this subject, our elastic thin-plate model takes into account the permanent deformation of the lithosphere caused by extension, that is, normal faulting and ductile deformation. As the presence of weak layers in the continental lower crust and a more detailed shape of the deformed lithosphere, including faults, are taken into account, the proposed new thin-plate model also differs from the finite element model utilized by Cloetingh et al. [1985]. Our model is based on the elastic thin-plate approach and does not incorporate viscoelasticity because, in general, flexural subsidence profiles for rifted lithospheres predicted by elastic thin plate models are in good agreement with observations [e.g., Watts et al., 1982]. Furthermore, the values for the extra parameters in viscoelastic thin-plate models operating on a million year timescale are not well constrained [Watts et al., 1982]. However, if required, the approach outlined in this paper can be applied to extend existing viscoelastic thin-plate models which include the effect of in-plane horizontal stresses [e.g., Lambeck, 1983].

In the first part of the paper, we outline the major differences between existing models and our new model. Subsequently, we review lithospheric flexure, with special emphasis on decoupled flexure of continental lithosphere. Next, our new thin-plate model for in-plane stress-induced differential motions is outlined. Particular attention is paid to the specific response of faulted lithosphere and its effect on the sedimentary record. This is followed by a comparison of the elastic thin-plate solution and the results of finite element modeling. In-plane stress-induced differential subsidence patterns are presented for several fault and basin configurations. Finally, we discuss the relationship between observed differential subsidence of the Pannonian and North Sea basins and the Norwegian margin and the major fault systems occurring in their basement.

2. Previous Research

The effect of horizontal in-plane stress variations on the subsidence of rifted sedimentary basins was first investigated by Cloetingh et al. [1985], who applied a two-dimensional elastic finite element model. The behavior of two different model sets was studied. One set of models employed a uniform elastic layer, whose thickness was increased with age according to a square root function. The second set of models adopted a laterally variable elastic thickness, representing the basinward decrease of the effective elastic thickness inferred for passive margins [e.g., Caldwell and Turcotte, 1979; Watts et al., 1982]. The results for the uniform elastic thickness model showed that an increase of compressional in-plane stress (stress times elastic plate thickness) causes uplift of the basin flank, contemporaneous with subsidence of the basin center, thus inducing a relative sea level change with different sign and magnitude across the basin system. In this model result, the magnitude of differential subsidence is controlled by the thickness of the elastic plate and the distribution and magnitude of the applied sediment load. The finite element models employing a laterally variable elastic plate thickness predicted opposite differential vertical motions for the same stress change, even in the absence of a sediment load. This completely different result was attributed to the lateral variation in plate thickness by Cloetingh et al. [1985]. However, as will be shown below, the geometry of the elastic plate is the cause for this response.

According to the thin-plate theory, a horizontal in-plane force applied to a predeformed plate gives rise to moments, causing additional flexural bending of the plate (Figure 1) [Timoshenko and Woinowsky-Krieger, 1959]. The moment equals the product of the in-plane force and the slope of the midplane of the elastic plate. The slope of the midplane is caused by preexisting deformation and by bending due to the in-plane force itself. The preexisting deformation can be permanent, flexural, or a combination of both. However, what is the preexisting deflection of the midplane of a thin plate representing the lithosphere of a rifted basin? This difficulty is illustrated by the differences in published elastic thin-plate models. Karner [1986] applies the in-plane force change to the preexisting deflections of the sediment-basement interface and the Mohorovicic discontinuity and adds the two resulting deflections to obtain a net deflection of the lithosphere. His justification for this procedure is that these two surfaces represent density interfaces, which can potentially be affected by lithospheric in-plane stress. Subsequently published thin-plate models [Cloetingh, 1988; Kooi and Cloetingh, 1992; Karner et al., 1993] basically all assume that the preexisting deformation of the midplane is caused by the flexural bending of the lithosphere due to thermal, sediment and necking loads (Figures 2a and 2b). Permanent deformation of the lithosphere due to normal faulting in the upper crust and extension-driven ductile flow in the lower crust and mantle, which also deform the midplane, is not included. Therefore, as these models do not incorporate all the preexisting deformation of the lithosphere, the predictions of these models cannot be correct. For basin modeling purposes, the models of Kooi and Cloetingh [1992] and Karner et al. [1993] improve the older kinematic basin evolution models by taking into account that, because of a deep level of necking during extension, the lithosphere can be
3. Effective Elastic Thickness and Rheology of Oceanic and Continental Lithosphere

For an elastic thin-plate, fiber stresses resulting from bending are zero at the plate center and increase linearly in magnitude outward according to a gradient defined by the Young's modulus, Poisson's ratio, and the curvature of the plate (Figure 4a) [Timoshenko and Woinowsky-Krieger, 1959]. The bending moment in the plate is obtained from the integral with depth of the fiber stresses with respect to the midplane of the plate. The flexural rigidity \( D \) of an elastic plate is defined as the ratio of the bending moment and curvature of the bent plate. For an elastic thin plate, the flexural rigidity is a function of the plate's thickness \( t \) and its elastic properties [Timoshenko and Woinowsky-Krieger, 1959]:

\[
D = \frac{ET^3}{12(1-v^2)}
\]

where \( E \) is Young's modulus and \( v \) is Poisson's ratio.

However, fiber stresses of a lithospheric plate are limited by its yield strength [McNutt and Menard, 1982; McNutt et al., 1988; Ranalli, 1994; Burov and Diament, 1995], which varies with depth according to Byerlee's law for brittle failure and ductile creep laws [Kirby, 1983; Carter and Tsenn, 1987].

![Figure 3](image)

![Figure 4](image)
The rigidity of a decoupled system is much smaller than the rigidity of a welded system of the same thickness bent to the same curvature. Therefore the decoupling model can explain that the $T_c$ of the lithosphere is generally larger than the theoretical $T_c$ of the crust alone but is smaller than the sum of crust and subcrustal $T_c$ [Burov and Diament, 1995, 1996].

The rheological state of the lithosphere controls whether or not the crust and the subcrustal lithospheric mantle are mechanically decoupled; it depends mainly on the thickness of the crust and the thermal state of the lithosphere. For old (cold) lithosphere, the critical thickness is about 35 km, which coincides with the average crustal thickness. This suggests that most of the continents are in a decoupled mode [Burov and Diament, 1996]. About 75% of observed continental $T_c$ can be explained by flexural decoupling [Burov and Diament, 1995, 1996; Cloetingh and Burov, 1996].

4. New Stress-Induced Differential Subsidence Model

The preexisting deformation of an elastic thin plate affects the response of the plate to a change in the magnitude of the in-plane force [Timoshenko and Woinowsky-Krieger, 1959]. For a single elastic plate an increase in the amount of the far-field in-plane force amplifies the existing bend $w_c$ whereas a decrease of force will flatten it. The flexural response is given by [Timoshenko and Woinowsky-Krieger, 1959]

$$\frac{\partial^2 (D\partial^2 w)}{\partial x^4} + F \frac{\partial^2 (w + w_c)}{\partial x^4} + w(\rho_m - \rho_{III})g = 0$$

where $w$ is force-induced deflection, $F$ is in-plane force change, $D$ is rigidity, $\rho_m$ is density of asthenosphere, $\rho_{III}$ is density of material filling in the depression (sediment and water) and $g$ is gravitational acceleration. In the derivation of (2), $w_c$ represents the preexisting shape of the midplane of the thin plate before the force is applied [Timoshenko and Woinowsky-Krieger, 1959]. The preexisting deformation can be permanent, elastic, or a combination of both. The total deflection of the midplane is given by $(w + w_c)$.

In the decoupling model, however, the lithosphere consists of two or more superimposed elastic thin plates. This can be accounted for in the flexure equation by noting that each elastic plate carries part of the total in-plane force applied to the lithosphere [Cloetingh and Burov, 1996]. Each of these separate forces acts on a deformed midplane, with the shape of the preexisting deformations differing between the individual plates forming the composite beam. For example, the stress-induced deflection for a continental lithosphere consisting of two elastic plates, one residing in the upper crust and one residing in the upper subcrustal mantle, is described by the following equation:

$$\frac{\partial^2 (D\partial^2 w)}{\partial x^4} + F_1 \frac{\partial^2 (w + w_1)}{\partial x^4} + F_m \frac{\partial^2 (w + w_m)}{\partial x^4} + w(\rho_m - \rho_{III})g = 0$$

where $D$ is rigidity of decoupled elastic plate system, $F_1$ is in-plane force change in the crustal layer, $F_m$ is in-plane force change in the lithospheric mantle, $w_1$ is preexisting deflection of crustal midplane, $w_m$ is preexisting deflection of subcrustal mantle midplane and $w$ is deflection of the lithosphere caused
by the in-plane force. Assuming an elastic response and no
stress relaxation in the fibers, \( F_C \) and \( F_m \) can be related to the
total force \( F \) acting on the lithosphere and the thicknesses of
the upper crustal and subcrustal competent layers:
\[
F_C = \frac{h_c F}{h_c + h_m} \quad F_m = \frac{h_m F}{h_c + h_m}
\]
where \( h_c \) is upper crustal competent layer thickness and \( h_m \) is
subcrustal fiber thickness.

An approximation for the shape of the preexisting
deformation of the crustal elastic plate is given by
the basement topography. However, the basement topography
must be corrected for erosion which is likely to take place
during the synrift and early postrift phase of the basin,
particularly in the area of the rift shoulders. As the midplane is
located about halfway between the upper and lower
boundary of the upper crustal competent layer, it is displaced
downward by half the magnitude of erosion (i.e., the bottom
of the layer is not modified by erosion).

\( w_c = \) basement topography + erosion amount / 2

The topography of the Moho discontinuity is probably the
best approximation for the shape of the preexisting
deformation of the subcrustal flexural midplane, as it is
located close to the strongest part of the mantle lithosphere
predicted by rheological modeling using yield-strength
envelopes [e.g., Cloetingh and Burov, 1996]. For basin
modeling purposes, both the Moho and the basement
landscape have the advantage that they can be observed and
mapped on deep-seismic profiles.

In the two-layer model, the flexural effect of a change in the
magnitude of an in-plane force depends on the curvature
of the midplanes of the lithospheric competent layers. As
outlined above, those parts of the continental lithosphere
contributing to its flexural strength are mainly located in the
upper crust and the uppermost part of the subcrustal mantle.
During extension processes, these parts are permanently
deformed by extensional brittle faulting and ductile flow.
As demonstrated by the results of deep seismic profiling
[Holliger and Klemperer, 1990; Bois, 1993; Posgay et al.,
1996], dynamic models for lithosphere extension [Zuber and
Parmentier, 1986; Braun and Beaumont, 1989b; Bassi et al.,
1993] and tectonostratigraphic forward modeling [Cloetingh
et al., 1995], the upper crust is generally displaced downward,
and the lithospheric mantle is displaced upward because of
lithospheric extension. Therefore the preexisting shapes of
the upper crustal and subcrustal fibers are opposite, as is the
contribution of these fibers to the total in-plane force induced
deflection. However, on deep seismic profiles of rifted
sedimentary basins, the Moho is relatively smooth compared
to the basement topography [Holliger and Klemperer, 1990;
Bois, 1993], although subcrustal faults can also occur [Flack
et al., 1990; Posgay et al., 1996]. During the postrift phase of
basin subsidence the Moho relief is reducing because of
thermal subsidence. Moreover, lower crustal flow, occurring
on a large timescale, may contribute to flattening of the Moho
during the postrift stage [Kaufman and Ronden, 1994; Ter
Voorde et al., 1998]. If the Moho topography has a negligible
curvature, then the preexisting shape of the midplane of the
subcrustal competent layer contributes little to the differential
vertical motions of the lithosphere upon the buildup of in-
plane force. In contrast, the shape of the sediment-basement
interface is generally highly irregular. This shape is an
approximation for the geometry of the upper crustal
competent layer. Therefore total basinal in-plane force-
induced deflections should be largely determined by the
effects of the geometry of the midplane of the upper crustal
competent layer. As a result, an increase in compressive in-
plane force will in most rifted basins cause a downward
deflection of the center of the rifted sedimentary basin,
because the preexisting upper crustal midplane geometry is
bent downward.

As can be observed on seismic profiles and in outcrops,
the upper crustal part of the lithosphere is permanently deformed
by discrete faults, causing a rugged basement topography. In fact,
in rifted basins most of the deformation of the upper crustal
competent layer results from faulting. Therefore, at
first sight, the curvature of the upper crustal flexural midplane
locally attains extremely high values, particularly along the
rift margins which are marked by large border faults with
throws of a few kilometers. However, owing to the presence
of faults, the upper crustal midplane is discontinuous, and
therefore the curvature of the midplane is not defined at the
position of faults. A new theory is required in order to resolve
this problem.

4.1. Faulted Lithosphere and Stress-Induced Deflection

A fault in the upper crust causes an offset in the upper

crustal competent layer. For vertical loading, the faulted
elastic plate behaves like an equally thick continuous plate
[e.g., Kuszmir et al., 1991], because the loads are acting
perpendicular to the plate. As faults and fractures are
widespread in the lithosphere, this is a basic assumption for
all models dealing with lithospheric flexure in response to
vertical loading. However, for horizontal loading the vertical
offset causes an additional moment at the fault, because the
in-plane stress is transmitted through the overlap zone of the
two parts of the plate which are separated by the fault. The
overlap zone, located at the fault, is narrower than the plate,
leading to a slight concentration of in-plane stress. The
horizontal in-plane stress profile in the overlap zone is not in
equilibrium with respect to the midplanes of the two
subplates, that is, at the overlap zone the integral of in-plane
stress with respect to the midplanes of the subplates is
nonzero. This causes a moment at the fault (Figure 5).

Seismotectonic and deep seismic reflection data suggest
that fundamental basement faults in rifted sedimentary basins
are planar and are restricted to the brittle, cold, topmost part
of the lithosphere, corresponding to the upper crust [King
et al., 1988; Jackson and White, 1989; Kuszmir et al., 1991;
Roberts and Yielding, 1991; Van Wees et al., 1996].
Therefore, in the model we constructed a vertical fault that is
assumed to cut the entire upper crustal competent layer. The
results of finite element modeling, presented in section 4.2,
demonstrate that the assumption of a vertical fault plane does
not significantly influence the prediction as long as the fault
dip is higher than 63°. Fundamental planar basement faults
must have the largest effect on stress induced deflections, as
they cause most of the deformation of the upper crustal
competent layer in rifted basins. However, elastic solutions
for faults not cutting the entire upper crust also exist [Savage
and Guohua, 1985] and possibly can be extended to include variations in in-plane force levels.

The farfiield in-plane force carried by a faulted thin plate is constant. Using this balance of in-plane in-plane force and the geometry of the faulted plate, it can be shown that the total in-plane force-induced moment at the position of a fault equals magnitude of the in-plane force times fault offset, \( F \alpha \) [Van Balen and Podladchikov, 1998]. Using general solutions for the thin-plate equation [e.g., Hetényi, 1946], the flexural profile caused by the in-plane force-induced moment can be found analytically for several scenarios, including variable rigidity [Van Balen and Podladchikov, 1998]. For a constant rigidity the elastic flexural profile due to the moment is described by

\[
w = \frac{F \alpha}{4D \sin(\beta)} x \sin(x \beta) e^{-x} x \alpha
\]

where \( w \) is deflection, \( D \) is flexural rigidity, \( F \) is in-plane force (positive for compression), \( \alpha \) is fault displacement, \( x \) and \( \beta \) are flexural parameters, \( \Delta p \) is density difference between infilling and compensating material, \( x = 0 \) corresponds to the fault position, and \( g \) is gravitational acceleration. The value of \( x \) is positive if the downthrown block is on the left side of the fault; otherwise, it is negative. Typical profiles predicted by this equation are presented in section 4.2.

4.2. Finite Element Modeling

In this section, the flexural profile predicted by (4) is compared to elastic finite element solutions. Two sets of models are analyzed: the first set employs a 10 km thick elastic plate, and the second set has a plate thickness of 5 km. These values for the elastic thickness are warranted by \( T \), estimates from the shapes of deflections around normal faults in rift settings [King et al., 1988; Kuszniir et al., 1991; Roberts and Yielding, 1991]. Rifting processes occur at the same timescale as in-plane stress level changes, that is, a few Myr (in fact, rifting is caused by an in-plane stress magnitude which exceeds the tensile strength of upper crustal lithosphere). Furthermore, flexural bending due to normal faulting occurs at the same spatial scale as fault-moment bending. Therefore we assume that both normal faulting during rifting and flexural bending due to fault moments are controlled by the same \( T \) (see also section 4.4).

All models apply 100 MPa far-field compression and a fault throw of +1 km, which is within the limits defined by the yield strength of the lithosphere [Kuszniir and Park, 1984; Savage and Guohua, 1985] and values deduced from studies on folding of oceanic [Cloetingh and Wortel, 1985; Beekman et al., 1996] and continental lithosphere [Lambeck, 1983; Cloetingh and Burov, 1996]. Apart from the intraplate stress, applied boundary conditions are zero vertical displacement at the lateral boundaries of the model (Figure 6). An array of springs at the bottom of the plate simulates the buoyancy conditions. The upper side of the model is a free surface. In order to reduce the effect of the lateral boundary conditions, the models have a total length of 1500 km. For some models, the plate is not allowed to thicken under the application of compressive stress thus mimicking the thin-plate assumption made in the derivation of the general elastic flexure equation [Timoshenko and Woinowsky-Krieger, 1959]. The adopted values for Young's modulus, Poisson's ratio and the density contrast between compensating materials are the same for the analytical and finite element models, that is, 7x10^10 Pa, 0.25, and 9400 kg m^-3, respectively. The applied density contrast is based on the density of fresh, surficial, hardly compacted sediments with a density of about 1800 kg m^-3 and a lower crustal density of 2700 kg m^-3.

For the 10 km thick model the analytical solution is compared to finite element results for models employing 1 km by 1 km and 2 km by 2 km quadratic finite elements, respectively (Figure 7a). Both numerical models predict the same result, indicating that the element size is not influencing the prediction. The difference between the analytical result and numerical solutions is up to 10% at the maxima of the flexural profile. The finite element model in which the plate is not allowed to thicken predicts a flexural profile which nearly coincides with the analytical prediction. For the 5 km thick plate the analytical solution is compared to finite element models with 1 km by 1 km and 2 km by 1 km elements, respectively (Figure 7b). Both these models predict the same deflection, indicating again that the element size does not influence the predictions. The maximum difference between numerical and analytical solutions is now up to 14%. The finite element model without plate thickening deviates...
4.3. Predicted Flexural Profiles for Multiple Fault Moments and Variable Rigidity

In-plane force-induced flexural profiles for fault systems can be obtained by applying the method of elastic superposition (Hetényi, 1946), that is, by adding up the flexural profiles for the single fault solution (Figure 8a). The flexural profile for a moment occurring at a single normal fault, representing a half graben, was obtained by applying a 5 km $T_e$, a fault throw of 2 km and an in-plane compression of 100 MPa carried by the whole plate (resulting in an in-plane force of $5 \times 10^{11}$ N m$^{-1}$). As shown by (4), the magnitude of the differential subsidence caused by the fault moment scales linearly with applied in-plane force and fault displacement. Therefore the flexural profiles presented below can also result from a combination of larger fault throws with less in-plane stress and vice versa. The elastic parameters have the same values as in the previous models. The predicted flexural profile for the half graben is characterized by a maximum uplift of about 75 m at the footwall and an equal amount of subsidence at the hangingwall (Figure 8a). The two maxima are separated by 50 km. Therefore the slope of the differential subsidence profile is at its steepest position equal to 0.17°. The differential subsidence profile predicted for a system consisting of two master normal faults dipping in opposite directions and spaced 50 km apart, delineating a simple rift, shows almost 150 m subsidence in the graben and 75 m uplift at the flanks. The larger amount of subsidence in the graben is caused by positive interference of the two flexural profiles induced by the fault moments. The maximum slope in this profile is 0.23°. A model involving a set of five master normal faults spaced 25 km apart and having throws in the same direction predicts a flexural profile with maxima of 140 m (Figure 8c). Finally, a model combining two opposing sets of five master faults predicts a maximum subsidence in the graben of about 210 m and a maximum uplift at the flanks of 170 m (Figure 8d). The maximum slope in this profile is 0.17°. The resulting profiles show that closely spaced multiple-fault systems generate positive interference, causing enhanced compression-induced subsidence at a scale larger than the individual tilted fault block scale, that is, basin-wide.
Figure 8. In-plane stress-induced differential subsidence profiles (continuous line), or relative sea level change, for different fault combinations. Faults have throws of 2 km. The elastic plate has a thickness of 5 km and is subjected to a 100 MPa compressive stress. Dashed lines indicate the faulted basement profile (vertical scale has to be multiplied by 10). (a) Resulting profile for a single fault. The compression induces 75 m uplift of the footwall and an equal amount of subsidence of the hanging wall. This profile is used to construct the predictions for the other scenarios by applying elastic superposition. (b) Simple rift with two master faults spaced 50 km apart. Predicted footwall uplift (flank) is about 75 m; stress-induced subsidence in the graben center equals 150 m. (c) Five faults with displacement in the same direction, spaced 25 km apart. The maximum uplift and subsidence are both about 140 m. The resulting differential subsidence pattern shows an almost linear trend across the faults, with a slope of 0.16°. (d) Results for two opposing sets of five faults. Within a set, faults are spaced 25 km apart and have the same throw direction. The two sets are spaced 50 km apart. The maximum of the predicted subsidence exceeds 210 m, whereas maximum uplift is about 170 m. The trends of differential subsidence across the fault sets are almost linear, with slopes of 0.17°.

4.4. Implementation of Fault Moment Solution

Although it is the most important part, the contribution of fault-related moments to the differential subsidence in a rifted basin is only part of the total response of the rifted basin to changing far-field stresses. The shape of the crustal midplane without faults and in some exceptional cases the shape of the mantle midplane (see below) also contribute. The total, basin-wide differential subsidence pattern can be obtained by superposition of the components contributing to it. The shape of the upper crustal elastic layer without the faults is found when the fault throws are reversed. After the fault offsets are
restored, the discontinuities in the basement profile are eliminated and therefore the resulting shape of the upper crustal elastic plate is differentiable. The flexural subsidence profile due to a changing in-plane force can be found by using a standard finite difference method.

For oceanic, thick, and hot continental lithospheres, (2) should be used in combination with the fault solution (4), because, owing to the rheological state, their flexure can be effectively described by one elastic plate [Cloetingh and Burov, 1996]. Stress-induced deflection for an average continental lithosphere should be calculated using (3) in conjunction with the fault solution (4). Flexure of a thin continental lithosphere underlain by normal or cold mantle is dominated by the mantle rheology. Therefore, in this exceptional case, the single elastic plate equation (2) should be used, and fault moment contributions can be neglected.

The $\mathbf{T}_e$ controlling the fault moments can be substantially less than the $\mathbf{T}_e$ of (decoupled) lithosphere. Values of $\mathbf{T}_e$ estimates from the shapes of deflections resulting from normal faulting are about 5-10 km [King et al., 1988; Kuszniir et al., 1991; Roberts and Yielding, 1991]. The timescales of normal faulting (e.g., rifting) and fault moments are the same, that is a few million years. Additionally, they have also the same spatial scale. Therefore they should be governed by the same $\mathbf{T}_e$. The $\mathbf{T}_e$ of large-scale deflected, (decoupled) rifted continental lithosphere is in the range of 10 - 20 km [Watts et al., 1987; Burov and Diament, 1995]. The low value of $\mathbf{T}_e$ for faulting and fault moments can be explained by a combination of two effects: a reduction of $\mathbf{T}_e$ around faults because of relaxation of fiber stress [Kuszniir et al., 1991], and the effect of the scale of the deflection or, equivalently, by the difference in load distribution [Ter Voorde et al., 1998]. Sediment, thermal, and necking loads typically have a basin-wide distribution. This causes a relatively minor pressure gradient on the lower crustal low-viscosity channel, inducing negligible lower crustal flow on a 10-100 My timescale. Normal faulting and fault moments, however, produce more localized loads and therefore cause higher pressure gradients in the lower crust. Therefore local lower crustal flow may provide flexural compensation for normal faulting and fault moments. The $\mathbf{T}_e$ for faulting and fault-moments is thus controlled by the effective elastic thickness of the upper crust alone. Additionally, the relaxation of fiber stress in the vicinity of faults may cause a reduction of $\mathbf{T}_e$ compared to the upper crustal value [Kuszniir et al., 1991].

4.5. Predictions for Total Stress-Induced Deflection

Applying the numerical procedure outlined in section 4.3, we have computed stress-induced differential subsidence and uplift of a decoupled continental lithosphere for different rifted basin scenarios. In all examples, the total $\mathbf{T}_e$ of the decoupled lithosphere is 15 km, and the $\mathbf{T}_e$ for the fault moments is 5 km. The adopted in-plane stress level change is 100 MPa compression. Elastic parameters have the same values as in previous examples. The density difference for lower crustal compensation is taken as 700 kg m$^{-3}$, asthenospheric compensation is calculated adopting a density difference of 1300 kg m$^{-3}$. Two opposing sets of five faults are located in the upper crust, each fault has a throw of 2 km.

Two types of models are analyzed, one set has a flat Moho discontinuity (Figure 9a), while the other set adopts a slight Moho uplift (Figure 9b). The models with a flat Moho represent a basin originating from a deep level of necking during extension. The models with a Moho uplift represent a basin which developed with an intermediate depth of necking, located in the lower crust. Each model set contains three configurations for the total $\mathbf{T}_e$ distribution. As the $\mathbf{T}_e$ results from decoupling, it can be located mainly in the upper crust or mainly in the subcrustal mantle, or it can be about equally distributed between the mechanically strong layers, resulting in the same total $\mathbf{T}_e$ according to (1). In the models, combinations of upper crustal and subcrustal competent layer thicknesses of 5 km and 14.8 km, 11.9 km and 11.9 km, and 14.8 and 5 km are applied.

The result for the models adopting a flat Moho show 100 to 200 m flank uplift and 180 to 240 m basin center subsidence (Figure 9a). The model with the largest upper crustal competent layer thickness (14.8 km) predicts the largest amplitude of deflection. The models adopting a slight Moho uplift predict 150 to 200 m flank uplift and 150 to 190 m basin center subsidence, also depending on the competent layer thickness distribution (Figure 9b).

For the models adopting thicknesses of 5 and 14.8 km for the upper crustal and subcrustal competent layers, respectively, the separate contributions from the faults, the geometry of upper crust without the faults, and the deformed shape of the mantle to the total stress-induced deflection is shown in Figure 9c. In this model, upon compression, the subcrustal competent layer causes most of the flank uplift, whereas the faults induce the basin center subsidence.

The results demonstrate that with increasingly deeper levels of lithospheric necking during rifting the stress-induced differential subsidence amplitude increases. Furthermore, the total amount of stress-induced subsidence and uplift depends on the thickness of the upper crustal competent layer, because the total in-plane force acting in this layer is the product of in-plane stress and the thickness of the layer. Finally, the models representing a basin originating from an intermediate level of necking during extension predict flank uplift which is greater than the basin center subsidence, leading to a considerable narrowing of the basin. The flank uplift is mainly caused by stress-induced deflection of the subcrustal upper mantle.

5. Examples of Stress-Induced Subsidence and Uplift in Rifted Basins Related to Faults

5.1. Southern North Sea Basin

During the Tertiary, the overall development of the North Sea basin is characterized by subsidence rates in accordance with the post rift development following Triassic to Early Cretaceous rifting events [Ziegler, 1990; Kooi et al., 1991]. During late Pliocene-Quaternary times unusually rapid subsidence and sedimentation occurred in the southern part of the North Sea basin, leading to a Quaternary sediment thickness of up to 1 km (Figure 10). Since most sediments were deposited in a shallow-water environment, this part of the North Sea basin must have experienced an extremely high tectonic subsidence rate during the Quaternary [Kooi et al., 1991].

Earthquake focal mechanism solutions and borehole breakout data indicate that the present-day stress field in the
Figure 9. Predicted differential subsidence for rifted basin geometries originating from deep and shallow levels of lithospheric necking during extension. For each geometry, the total \( T \), of 15 km is distributed over the upper crustal and subcrustal competent layers as combinations of 5 km and 14.8 km, 11.9 km and 11.9 km, and 14.8 and 5 km. The applied in-plane stress is 100 MPa. (a) The results for the models representing deep necking show 100 to 200 m flank uplift and 180 to 240 basin center subsidence. The model with the largest upper crustal competent layer thickness predicts the largest amplitude of stress-induced deflection. (b) The models adopting a slight Moho uplift, representing an intermediate level of necking, predict 150 to 200 m flank uplift and 150 to 190 m basin center subsidence, depending on the thickness of the upper crustal competent layer. (c) The separate contributions of upper crustal faults, geometry of upper crust without faults, and shape of the subcrustal lithosphere to the total deflection for the model adopting a slight Moho uplift and thicknesses for the upper crustal and subcrustal competent layers of 5 and 14.8 km, respectively. The faults contribute most to the predicted subsidence in the basin center, whereas the subcrustal stress induced deflection causes the largest part of the basin flank uplift. See text for further discussion.
Figure 10. (a) Contour map of the Quaternary thickness distribution in the southern part of the North Sea basin [Kooi et al., 1991; Van den Berg, 1994]. Axial labels are in hundreds of meters. The maximum thickness locally exceeds 1 km. (b) Main Mesozoic fault zones and rift systems in the southern North Sea basin [Ziegler, 1990; Kooi et al., 1991; Van den Berg, 1994]. Abbreviations are as follows: MF, Moray Firth; VG, Viking Graben; CG, Central Graben; DCG, Dutch Central Graben; SP, Sole Pit basin; BF, Broad Fourceans basin; HG, Horn Graben. A good correlation exists between the N-S orientated faults and the local maxima (plus sign) and minima (minus sign) of Quaternary subsidence. Hanging walls correspond to maxima; footwalls correspond to minima. See text for discussion.

The southern part of the North Sea basin is characterized by a maximum compressive stress orientated NW-SE [Ahorn, 1975; Illies, 1975; Müller et al., 1992; Lindholm et al., 1995]. The NW-SE compression is caused by a combination of Alpine collisional forces and North Atlantic ridge push. The effect of the stress field on Quaternary subsidence was investigated by Kooi et al. [1991]. Their results indicate that the anomalous subsidence can be partly explained by a combination of a 300 MPa far field compressive stress and a decrease of $T_c$ from 20 km at the rims of the basin to 1 - 2 km below the Quaternary centers of maximum deposition (giving rise to an unrealistic local in-plane stress of about 3000 MPa). The fit was improved by adopting a decrease of paleo-water depth during the Quaternary. However, their model requires a high amount of far-field stress and a palaeo-water depth which is not in accordance with observations. Therefore Kooi et al. [1991] favor a model based on hypothetical local extension below the Quaternary centers of maximum deposition because of pull-apart tectonics related to strike-slip faulting, for which no evidence exists. A three-dimensional forward model of Quaternary stress-induced subsidence by Van Wees and Cloetingh [1996], also excluding the permanent deformation of the lithosphere, utilizes a lower far-field in-plane stress level (150 MPa) compared to the two-dimensional flexural model of Kooi et al. [1991]. The 3-D flexure model can explain up to 70% of the total Quaternary thickness in the Dutch Central Graben (Figure 10) but fails to predict the remaining patterns of Quaternary deposition. Furthermore, their model also requires a substantial decrease of $T_c$ in the Dutch Central Graben area (from 15 km at the flanks to less than 5 km in the center). Therefore Van Wees and Cloetingh [1996] also suggest that other tectonic mechanisms, like pull-apart extension, have played a significant role in Quaternary subsidence of the southern North Sea basin. There is, however, no evidence for major, large-scale Quaternary faulting, and the relationship between the NW-SE directed maximum compressive stress directions and the geometry of the fault system requires that the N-S orientated centers of maximum deposition are transpressed by the proposed strike-slip motions and not pulled apart.

The thickness distribution of Quaternary sediments is characterized by a saucer shaped long-wavelength component and local elongated thickness maxima and minima (Figure 10a). The largest Quaternary centers of deposition in the southern North Sea basin, located in the Dutch Central Graben area, are on the downthrown sides of N-S striking Mesozoic normal faults (Figure 10b) [Kooi and Cloetingh, 1989; Van den Berg, 1994]. Local minima of Quaternary thickness are on the footwalls of these faults. The strike of the N-S normal faults has an angle of 45° to the maximum compressive stress direction. The NW-SW orientated faults, parallel to the maximum compressive stress direction, are not associated with Quaternary centers of deposition and...
thickness minima. Furthermore, in the Central Graben and Egersund Basin of the northern North Sea basin little or no increase in subsidence has occurred [Hail and White, 1994]; these basins strike parallel to the maximum compressive stress direction. The N-S striking faults have a component of the maximum compressive stress acting on them. Therefore, we suggest that the Quaternary centers of maximum deposition and thickness minima can be explained by compressive stress-induced subsidence related to moments acting at the position of the Mesozoic faults. Faults with throws of several kilometers (up to 7 km) are common in this part of the North Sea basin [Ziegler, 1990; Holliger and Klemperer, 1990]. As the differential subsidence caused by the moments scales linearly with the fault displacement and in-plane force, the results in Figure 8 can be used to estimate the contribution of the moments to the differential subsidence (a $T_s$ of about 5 km for the faulted upper crust of the North Sea basin is inferred by Roberts and Yielding [1991]). In the centers of maximum deposition, the thickness of Quaternary sediments is about 400 m more than the regional long-wavelength trend. This can, for example, be explained by a combination of fault throws of 5 km and a far-field stress of 100 MPa. The total Quaternary thickness distribution is then the sum of small-wavelength fault-related differential subsidence and the long-wavelength Quaternary distribution. The latter can be explained by the compressive stress response of the downflexed lithosphere caused by necking, sediment, and thermal loading.

5.2. Pannonian Basin

The Pannonian basin is a Neogene intramontane basin system bounded in the west by the Alps, in the north, east, and southeast by the Carpathians, and in the southwest by the Dinarides. The Pannonian basin originates from contemporaneous tectonic escape in the Alps, asthenospheric upwelling, and subduction rollback along the Carpathian front [Ratschbacher et al., 1991; Horváth, 1993]. Because of changes in plate motions, the stress field in the Pannonian basin changed to compression during late Pliocene [Csontos et al., 1991; Müller et al., 1992]. The maximum horizontal stress direction in the southern part of the basin is roughly SW-NE oriented and in the northern part coincides with the European NW-SE trend [Müller et al., 1992]. During late Pliocene-Quaternary times, the central part of the basin system experienced an acceleration of subsidence, whereas the external parts were uplifted, with the exception of the local subbasin centers in the southwestern and northwestern parts of the basin system (Figure 11a). Differential subsidence took place without major fault reactivation [Horváth and Cloetingh, 1996], indicating a flexural mechanism. Uplift in excess of 300 m occurred in the Trans-Danubian central range, whereas about 1500 m of sediment accumulated in the Makó Trough and Békés Basin during the same time interval. Using the previously published thin-plate model [i.e., Kooi and Cloetingh, 1992], the differential subsidence pattern can be explained by the buildup of the contemporaneous compressive stress field, adopting a 300 MPa in-plane stress, in combination with ongoing thermal subsidence in the center of the basin system [Horváth and Cloetingh, 1996; Van Balen et al., 1998]. A comparison of the differential subsidence pattern with the fault system in the basin shows a clear relationship (Figure 11b), evidencing that the faults exert a first-order control on the stress-induced differential subsidence. The subsidence and uplift in the Sava and Drava Troughs in the southwest and the Little Hungarian Plain (LHP) in the northwest are perfectly fault bounded. The faults bounding the Makó Trough and Békés Basin are overstepped by late Pliocene-Quaternary sediments, indicating a flexural rigidity which is less for the hanging walls compared to the footwalls [Van Balen and Podladchikov, 1998]. This can be explained by the high heat flow in the Makó Trough and Békés Basin [Posgay et al., 1996], causing weakening of the lithosphere.

5.3. Norwegian Margin

The morphology of Norway, in combination with the almost complete absence of post-Paleozoic sediments, evidences large-scale Tertiary uplift of western Fennoscandia. Two domal uplifts in southern and northern Norway are inferred from the present position of preserved planation surfaces [Riis and Fjeldskaar, 1992; Riis, 1996] and results from fission-track thermochronology [e.g., Rohrmann et al., 1995; Riis, 1996]. These uplifts are spatially separated by a low-lying region. The onshore domes correspond to narrow offshore shelves, whereas the low-lying saddle in between them is associated with the broad shelf of the mid-Norwegian margin [Stuevold and Eldholm, 1996]. The domes are elongated and roughly parallel to the Atlantic coastline. Their maximum elevation is about 2 km. The flexural nature of the domal uplifts is evidenced by their shape and the absence of major fault activity associated with their development. The observed asymmetry of the domes, consisting of a steep westward face and a more gentle slope eastward, is explained by amplified weathering on their seaward flank [Stuevold and Eldholm, 1996].

The domal uplift in southern Norway is bounded in the south and west by the North Sea basin and in the northwest by the Møre Basin and the mid-Norwegian margin (Vrång Basin) (Figure 12). These basins were formed in response to Permo-Triassic to Early Cretaceous rifting. A final Paleocene extension phase in the Møre Basin and the mid-Norwegian margin caused the opening of the North Atlantic Ocean. The structural, basinward dip of strata and deposition of large clastic wedges in the Norwegian offshore sedimentary basins record substantial Tertiary uplift and erosion of the mainland and subsidence in the basins [Stuevold et al., 1992; Stuevold and Eldholm, 1996; Riis, 1996].

The flexural uplift and subsidence of the Norwegian mainland and the surrounding basins is separated in time from the opening of the North Atlantic. Several mechanisms have been proposed, including convective/thermal processes, intraplate compression, rift shoulder uplift, glacial erosion, mantle plumes, and phase changes. However, none of these can satisfactorily explain the uplift and associated subsidence (for an overview see Stuevold and Eldholm [1996]). In addition, the exact timing of uplift is controversial; a late Oligocene to early Miocene time is suggested by Rohrmann et al. [1995] and Stuevold and Eldholm [1996]. Recently, using morphological evidence and offshore seismic reflection data, Riis [1996] proposed that the Tertiary uplift and subsidence
took place in two phases: a Paleogene phase and a Plio-Pleistocene phase, each with an equal amount of uplift. The hinge lines of the domal uplifts, separating uplifted from downwarped areas, coincide almost everywhere with major basement faults. Along the North Sea coast, the hinge line coincides with the Øygarden Fault Zone (Figure 12) [Riis, 1996], which consists of two major faults, each with a throw of more than 5 km toward the basin [Gabrielsen et al., 1990]. Northward, the hinge line is located at the position of the Møre-Trøndelag Fault Zone. On the mid-Norwegian margin, the hinge line coincides with the Bremstein and Revfallet fault complexes [Stuevold and Eldholm, 1996]. This correlation pattern continues farther northward [Riis, 1996]. Therefore it seems that the major fault zones exerted a first-order control on the distribution of uplift and subsidence. The stress system of the Norwegian mainland and adjacent basins is comparable to that of the North Sea area [Bungum et al., 1991; Müller et al., 1992; Lindholm et al., 1995]. Furthermore, if mantle upwelling has contributed to the domal uplift of the Norwegian mainland, it can explain part of the compressive stress distribution [e.g., Poliakov et al., 1993], and the total uplift may be the result of a stress-induced amplification of the mantle-caused uplift.

6. Discussion

In a recent paper, Karner et al. [1993] stated that the development of global or regional unconformities by in-plane force variations is difficult to accept for the following reasons: (1) localized fault reactivation dominates the response of the lithosphere to applied forces, (2) the orientation and sign of the present-day stress field changes markedly across the same plate, (3) the magnitude of the in-plane force required to modify the stratigraphy of an evolving basin is extremely large, and (4) the flexural deformation is governed by factors that will give very different responses of the lithosphere for the same in-plane force. However, not every in-plane force change leads to active faulting; as long as the prevailing in-plane force is at a level lower than the integrated strength of the lithosphere underlying the basin (including weakness zones due to preexisting faults), flexural deformation of the basin dominates. For example, the latest Cretaceous inversion of the mid-Polish Trough is preceded by a flexural acceleration of subsidence. Possible analogues are the rapid Paleocene-Early Eocene phase of tectonic subsidence of the Sverdrup Basin in arctic Canada [Dadlez et al., 1995; Stephenson, 1996] and the acceleration of subsidence during the Paleocene in the central North Sea [Ziegler et al., 1995]. These examples show that prior to fault reactivation, flexural subsidence in response to the buildup of compressional stresses occurs in the basin center. Only if the in-plane force reaches a threshold defined by the strength of lithosphere [Kuszmir and Park, 1984], will fault reactivation take place. Therefore, if the in-plane force does not reach this threshold, the lithosphere will respond to changing in-plane forces by flexure. The in-plane force of $5 \times 10^{11} \text{N m}^2$ adopted in the models in this paper is within the limits defined by the yield strength of the lithosphere [Kuszmir and Park, 1984; Savage and Guo, 1985], and values deduced from studies on folding of oceanic [Cloetingh and Wortel, 1985; Beekman et al., 1996] and continental lithosphere [Lambeck, 1983; Cloetingh and Burov, 1996].
measured and calculated stresses (up to 300 MPa) can be explained by the ductile nature of the lower part of the crust which causes a redistribution of stresses applied to the whole lithosphere and their amplification in the more brittle parts of the lithosphere [Kuszniir and Park, 1984; Savage and Guohtua, 1985]. Furthermore, the thickness over which the in-plane force is distributed should be taken into account. For example, a ridge push force of 2.4x10^15 N m^-1 applied to 80 km thick lithosphere may produce 30 MPa, whereas the same force applied to 160 thick lithosphere gives only 15 MPa [Kuszniir and Park, 1984]. If the same in-plane force is applied to a 15 km thick strong part of the lithosphere, a stress of 160 MPa results.

The flexural response of rifted lithosphere to an in-plane force change is dominated by the faulted shape of the upper crustal competent layer and, in exceptional cases, also the subcrustal mantle fiber. In rifted basins, the large-scale shape of the faulted upper crustal flexural midplane is always downward bend. Therefore the response of rifted basins to an in-plane force change should always be the same: basin center subsidence and flank uplift. This confirms the original results of Cloetingh et al. [1985]. The elastic thin-plate models including necking of the lithosphere during extension [Kooi and Cloetingh, 1992; Karner et al., 1993] concluded that the deeper the level of necking is during extension the less stress-induced differential vertical motion there is during the postrift phase. Also, if the level of necking during rifting was deep enough, the flexural response to in-plane stress changes during the postrift stage would be reversed compared to a situation with a shallow level of necking during extension. However, a deeper level of the lithospheric necking causes an increased deformed geometry of the upper crustal competent layer. As a result, the post rift stress-induced vertical motions increase when the synrift lithospheric necking level is deeper.

Using the argument that stress-induced lithospheric deflections cause significant tilting of sedimentary strata, Karner et al. [1993] argued that parallel seismic reflectors are indicative for eustatic sea level changes. Our results demonstrate that the angle of tilting is so extremely low (less than 1°) that such tilting is hardly observable on seismic sections. Therefore parallelism of seismic reflectors is not necessarily an indication of a crustal control on onlap and offlap configurations.

After reversing the sign of the horizontal intraplate stress, our model predicts flexural uplift in the center of a rift basin in response to tension. The upward bending of the plate causes additional tensile fiber stresses in the center of the rift. These additional tensile stresses augment the far-field tensile stress and therefore will facilitate further extensional faulting in the rift. As a consequence, our model predicts that rifting will localize in preexisting rift grabens.

For the elastic stress-induced differential subsidence model, a decrease in tensile stress has the same effect as an increase in compressional stress. As shown by our results, an increase in compressional stress leads to uplift of a basin flank. Therefore the release of tensional stress when rifting ceases is a mechanism for flank uplift. Evidence from various passive margins demonstrates that erosion of a rift shoulder causes uplift of the nearby sedimentary basin, inducing stratigraphic onlap during the early phases of the postrift

Figure 12. Generalized Mesozoic Paleocene fault map of southern Norway and the mid-Norwegian margin [Gabrielsen et al., 1990; Blystad et al., 1995; Riss, 1996]. Abbreviations are as follows: ØFZ, Øygarden Fault Zone; MTFZ, Møre-Trøndelag Fault Zone; BFC, Bremsten Fault Complex; RFC, Revfallet Fault Complex; NSB, North Sea basin; MB, Møre basin; VB, Voring basin; and VMH, Voring marginal high. Ruled areas indicate the extent of the late Oligocene-Pliocene uplift [Riss, 1996; Stuevold and Eldholm, 1996]. The hinge line of the uplift coincides with major fault zones.
evolution of a rifted basin [Van Balen et al., 1995]. This can be explained by flexural compensation for erosion on a flank uplift [Van Balen et al., 1995], which amplifies the stress-induced uplift and the resulting relative sea level change.

Finally, application of the presented equation for fault moments is not restricted to faulted elastic plates. In fact, this equation can be used to study the effects of any steep lateral change of the geometry of a competent layer on in-plane stress-induced deflections. Such steep morphology may, for example, exist at ocean-continent boundaries.

7. Conclusions

Our new model for intraplate stress-induced subsidence takes into account flexure of a multilayered continental lithosphere and its structural configuration prior to the change in the in-plane stress regime. The preexisting deformation influences the response of a thin plate to a change in in-plane force (stress times plate thickness) considerably [Timoshenko and Woinowsky-Krieger, 1959]. Because the preexisting deformation of a rifted sedimentary basin has a concave geometry, the response of a basin to compression is subsidence in the basin center and uplift of the flanks. The preexisting geometry is not properly taken into account in published thin-plate models. The largest part of the preexisting deformation in rifted sedimentary basins is caused by faults in the upper crustal competent layer. The offset of the midplane of the upper crustal competent layer due to a fault causes an additional moment at the position of the fault. This moment causes subsidence on the downthrown side of the fault and uplift of the footwall. The amount of differential uplift depends on the vertical throw of the fault, the magnitude of the in-plane stress, the thickness of the upper crustal competent layer, and the degree of flexural coupling between the upper crustal and upper mantle competent layers. As shown in section 4, uplift and subsidence of several hundred meters are predicted for conservative values of these parameters in a basinal setting with multiple faults. Therefore the total stress-induced deflection, or relative sea level changes, of rifted basins can be of the order of 1 km in the case of a large in-plane stress level change (up to 300 MPa), large fault throws (several kilometers), and a strong upper crust in combination with a relatively weak subcrustal layer.

In-plane stress level changes caused by, for instance, lithospheric plate reorganizations [e.g., Cloetingh et al., 1989] can therefore have a substantial impact on relative sea level change in a rifted basin. As the lithospheric plates form an interlocking system, plate boundary reorganizations may cause a global change in in-plane stress level, and this may induce worldwide onlap and offlap events, interpreted as eustatic sea level changes [e.g., Haq et al., 1982].

Examples from the southern North Sea basin, Pannonian basin, and the Norwegian margin demonstrate that faults do, indeed, play a key role in the pattern of observed stress-induced differential subsidence of rifted sedimentary basins during their postrift evolution. The results of our new model confirm the original hypothesis of Cloetingh et al. [1985]: an increase in the level of compressive far-field in-plane stress causes flank uplift and basin center subsidence.

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S. A. P. I. Cloetingh, Y. Y. Podladchikov, and R. T. Van Balen, Faculteit der Aardwetenschappen, Vrije Universiteit, De Boelelaan 1085, 1081 HV Amsterdam, the Netherlands. (e-mail: cloeting@geo.vu.nl; podl@geo.vu.nl, balt@geo.vu.nl)

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