Intraplate Stresses and the Subsidence History of the Sirt Basin

Freek van der Meer 1 and Sierd Cloetingh 2

التضاغط داخل الألواح وتاريخ الإنخفاض لموض سرت

فيري د فان دير مير و سيرد كلوتينج

الخلاصة

لقد أوضح تحليل الينخفاض لعشرتين بدأ تيقوي قطع القاعدة الست الرئيسية بحوض سرت التertiary تاريخ إنخفاضي مستمر لكل قطعة وتم التعرف على أربعة مراحل واضحة من التطور التكتوني من خلال تقييم منحنى الانخفاض ودرجة التطور التكتوني الإدخالي في حوض سرت.

المرحلة الأولى (1) وبدأ الانخفاض في السينوباين ولاكن بدرجة بطيئة.

المرحلة الثانية (2) وإذا إنخلت المرحلة الأولى فجأة، خلال مرحلة الكامباني المتأخر وتبعها مباشرة مرحلة التصدع القطني الإختلافي خلال الماستريخي والباليوسين.

أما المرحلة الثالثة (3) فهي تي قي التطور التكتوني لموض سرت خلال الأوروبين المبكر والانخفاض خلال الأوروبين المتأخر.

في حين بدأت المرحلة الرابعة (4) خلال عصر الأوروبين والباليوسين المبكر وتميزت بقطرة من التطور التكتوني البطيء.

وتسجل هذه المرحلة واضحة طوراً أنخيذياً شاملاً بأحواض السرت، وهي تتميز مرحلة قبل الشد مع انخفاض تكتوني قليل (المرحلة الأولى)؛ المرحلة المصاحبة للشذ زرع بنية انخفاض تكتوني سريع مع تأسيس مرحلة التصدع الإختلافي (المرحلة الثانية)؛ مرحلة ما بعد الشذ والتي توضح ببطء في الانخفاض نتيجة التأثير الحربي (المرحلة الثالثة والرابعة).

ويمكن شرح الانخفاض الحربي (معالج الشذ) بالنموذج الحربي الميكانيكي مع الشذ السريع عند حدث 75 مليون سنة مضت إلا أن هذا النموذج لم يوفق في شرح بداية الانخفاض عند فترة الشذ وليس الانخفاض التكتوني عند مرحلة (مع) - وبعد الشذ - خلال تطور حوض سرت يجب أن تحدث مرحلة شد لفترة محددة من الينخفاض المتأخر حتى الباليوسين، وبعض من الخلافات التي أجريت مرحلة التأينة المتعلقة بالبركة المعقدة في الزمن الحالي وهذا له علاقة بالتأثيرات ما بين وحي أوروبا وأوقيف التي تتحرك في تطور حوض سرت.

Abstract. Subsidence analysis of 20 wells covering six major basement blocks in the Sirt Basin of Libya reveals a consistent subsidence history for each of the individual basement blocks. The evaluation of subsidence curves and tectonic subsidence rate plots outlines four distinct phases in the tectonic evolution of the Sirt Basin. Subsidence started in Cenomanian time with a phase characterized by relatively low subsidence rates (phase 1). This period was abruptly terminated in Late Campanian, followed by a phase of differential block faulting during the Maastrichtian-Palaeocene (phase 2). During the Eocene, the tectonic evolution of the Sirt Basin is characterized by distinct periods of uplift (Early Eocene) and subsidence (Late Eocene) (phase 3). The Oligocene and Early Miocene is characterized by a period of low and decelerating tectonic subsidence rates (phase 4). These distinct phases record the subsidence evolution typical for rift basins characterized by a pre-rift sequence with relatively low tectonic subsidence (phase I), a syn-rift phase characterized by rapid tectonic subsidence accomplished by crustal thinning and differential faulting (phase II), and a post-rift evolution which shows decelerating subsidence with a mainly thermal origin (phases III and IV). Thermo-mechanical modelling with simple stretching models using an instantaneous rifting event at 75 Ma.

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explains the observed post-rift thermal subsidence but fails to explain the initial active subsidence during rifting. A Late Cretaceous-Palaeocene stretching phase of finite duration is needed to explain the observed tectonic subsidence during the syn- and post-rift evolution of the Sirt Basin. Our analysis demonstrates the key importance of stress-induced vertical motions observed in the Eocene to recent subsidence record for the Sirt Basin, related to the interaction of the European and African plates controlling the evolution of the Sirt Basin.

INTRODUCTION

Quantitative modelling of sedimentary basins is an important tool for the study of mechanisms governing the development of basins. The influence of the thermo-mechanical properties of the lithosphere on models of basin evolution has been studied by various workers (e.g. Beaumont and Tankard, 1987). Geodynamic processes contributing to basin subsidence include stretching of the lithosphere during the rifting phase (McKenzie, 1978), post-rift cooling and thermal contraction of the lithosphere amplified by sediment loading (Sleep, 1971), and flexural bending in response to vertical loading (Beaumont, 1978). In addition, intraplate stresses, intrinsically linked to changes in plate motions and the associated plate-tectonic evolution, have also been shown to affect the subsidence record and stratigraphic evolution of sedimentary basins (Cloetingh et al., 1985; Cloetingh, 1988). For example, short-term fluctuations in the intraplate stress fields, superimposed on the tectonic subsidence of the basin, are capable of inducing vertical motions at basin flanks at a rate and magnitude equivalent to short-term (1–5 m.y.) apparent sea-level fluctuations which are frequently interpreted in terms of glacio-eustatic control (Vail et al., 1977; Haq et al., 1987). According to this model, increasing the level of horizontal intraplate compression leads to a magnification of the flexural peripheral bulge flanking the basin, producing uplift of the basin flanks and subsidence of the basin centre. This gives rise to a seaward migration of the shoreline and subsequent development of sedimentary offlap. For a horizontal tensional intraplate stress field, the flanks of the basin subside and the basin centre is uplifted. This results in a landward migration of the shoreline and an apparent rise in sea level initiating a phase of sedimentary onlap (Fig. 1).

Recently, considerable progress has been made in the study of the lithospheric stress field (Zoback et al., 1989). Numerical modelling (e.g. Cloetingh and Wortel, 1985; Zoback et al., 1993) has resulted in a better understanding of the causes of observed variations in stress level and direction in lithospheric plates and has demonstrated a causal relationship between processes at plate boundaries and deformation of the plates’ interiors.

Studies on the tectonic evolution of the North Sea Rift Basin (Kooi et al., 1989), demonstrated the key importance of stress-induced vertical motions related to the interaction of the Eurasian and African plates.

The Mesozoic-Tertiary tectonic evolution of the African plate is directly linked to the opening history of the Atlantic Ocean and the dynamics of Africa-Eurasia convergence (Guiraud et al., 1992). The present paper investigates these relationships focusing on the stratigraphic record of the North African Sirt Basin in the context of the African-Eurasian plate interaction.

Libya is situated on the Tethyan margin of the African shield, and covers a system of intra-cratonic basins. The Sirt Basin, bordered in the east by the Dakhla Basin of
Egypt, in the west by the Ḥün graben, in the south by the Al Kufrah Basin, and in the north by the Mediterranean Sea, is the youngest of these basins. It formed during the Early Cretaceous-Tertiary in response to crustal extension causing active subsidence resulting in the collapse of the Sirt Arch (Conant and Goudarzi, 1967; Massa and Delort, 1984; Sestini, 1984). This rifting activity can be related to dextral shear along the Gibraltar-Maghrebian South Anatolian shear zone forming a diffuse trans-tensional plate boundary between Africa and the Italo-Dinaride block (Burke and Dewey, 1974; Dewey et al., 1974; Ziegler, 1988). Gealey (1988) interpreted the regional stratigraphy of the Sirt Basin and concluded that rifting was initiated during the Middle Cretaceous as a result of extension along a broad transform zone extending throughout the western Mediterranean into the Atlantic Ocean. More recently, Guiraud and Maurin (1992) discussed the Sirt Basin in the framework of Cretaceous rifting in West and Central Africa. Their study suggests a Neocomian-Barremian onset of rifting for the southern Sirt Basin, during which predominantly E–W trending half-grabens developed, followed by a Late Aptian rifting phase generating a system of NW–SE trending grabens. According to Gumati and Kanes (1985) and Gumati and Nairn (1991), Sirt Basin subsidence reached a climax during the Palaeocene–Eocene corresponding to a period of major crustal extension.

The present study of the subsidence record of the Sirt Basin focuses on the structure of the basin and differential movement of separate basement blocks during the Early Tertiary. The following six major basement blocks are sampled by wells: the West, East and Middle Zallah blocks, the Az Zerah platform, the Marâdah trough, and the Zaltan platform. The position of the analyzed wells is listed in Table 1 and shown in Fig. 2.

We first outline the stratigraphy of the Sirt Basin, followed by a discussion on the results of quantitative subsidence analysis. The tectonic subsidence rate curves established for six major basement blocks in the Sirt Basin provide evidence for the occurrence of four tectonic phases during the evolution of the basin. Attention is focused on the relation between intraplate stresses and basin stratigraphy by comparing subsidence predicted by stretching models (McKenzie, 1978; Rodden and Keen, 1980) and observed tectonic subsidence. Finally, a scenario is presented, relating the observed timing of tectonic events to relative plate motions between Africa and Europe and to changes in intraplate stress levels. The results presented in this paper synthesize two earlier papers by the authors (van der Meer and Cloetingh, 1993a, b).

### Table 1. Summary of wells used in this study.

<table>
<thead>
<tr>
<th>No.</th>
<th>Well</th>
<th>Basement block</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Cl-78</td>
<td>West Zallah trough</td>
</tr>
<tr>
<td>2</td>
<td>H1A-57</td>
<td>West Zallah trough</td>
</tr>
<tr>
<td>3</td>
<td>001-11</td>
<td>Mid-Zallah trough</td>
</tr>
<tr>
<td>4</td>
<td>AA1-11</td>
<td>Mid-Zallah trough</td>
</tr>
<tr>
<td>5</td>
<td>A1NC29-C</td>
<td>Mid-Zallah trough</td>
</tr>
<tr>
<td>6</td>
<td>B1-57</td>
<td>East Zallah trough</td>
</tr>
<tr>
<td>7</td>
<td>X1-11</td>
<td>East Zallah trough</td>
</tr>
<tr>
<td>8</td>
<td>P1-32</td>
<td>East Zallah trough</td>
</tr>
<tr>
<td>9</td>
<td>GG1-11</td>
<td>East Zallah trough</td>
</tr>
<tr>
<td>10</td>
<td>J1-57</td>
<td>East Zallah trough</td>
</tr>
<tr>
<td>11</td>
<td>G1-47</td>
<td>Az Zahrah platform</td>
</tr>
<tr>
<td>12</td>
<td>C1-11</td>
<td>Az Zahrah platform</td>
</tr>
<tr>
<td>13</td>
<td>NN1-11</td>
<td>Az Zahrah platform</td>
</tr>
<tr>
<td>14</td>
<td>E1-11</td>
<td>Az Zahrah platform</td>
</tr>
<tr>
<td>15</td>
<td>BBB1-11</td>
<td>Az Zahrah platform</td>
</tr>
<tr>
<td>16</td>
<td>F1-13</td>
<td>Marâdah trough</td>
</tr>
<tr>
<td>17</td>
<td>A1-94</td>
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</tr>
<tr>
<td>18</td>
<td>Y1-13</td>
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</tr>
<tr>
<td>19</td>
<td>I11-6</td>
<td>Zaltan platform</td>
</tr>
<tr>
<td>20</td>
<td>FF1-6</td>
<td>Zaltan platform</td>
</tr>
</tbody>
</table>

### BASIN STRATIGRAPHY

The sedimentary fill of the Sirt Basin (Fig. 3) overlies Precambrian igneous and metamorphic basement rocks and consists of Lower Palaeozoic clastic sediments (e.g. Gargaf group; Hofra and Amal formations) and Late Mesozoic Nubian Sandstones (pre-rift sequence) which are unconformably overlain by Late Cretaceous-Early Tertiary marine sediments (syn-rift sequence). This sequence is overlain by an Eocene-Early Miocene post-rift sequence characterized by a continental facies to the south and a shallow marine facies in the northern parts of the Sirt Basin.

Facies relationships within the Nubian Sandstone are complex due to local topographic variation at the time of its deposition (Pomeryol, 1968; Klitzsch and Squyres, 1990). A dominant southward change from sandstone-limestone-shale alternations to a continental facies was recognized by Barr (1968) and Barr and Berggren (1980).

Active rifting accompanied by volcanism commenced during the Late Cretaceous (Guiraud et al., 1992) resulting in the accumulation of shale deposits in the troughs and shallow marine carbonates on the highs (Bahî, Eiel, Rachmat formations and Sirt Shales). During Maastrichtian the depositional environment shallowed (Khalash Formation). This event was followed by a transgression and subsequent deepening of the basin at the onset of the Palaeocene (Barr and Berggren, 1980). A thick section of marine sediments was deposited in
the rapidly subsiding troughs (Hagfa and Beda formations). Carbonate build-ups developed on the platform areas with, locally, pinnacle reefs in basin areas (Terry and Williams, 1969). Toward the close of the Early Palaeocene, basin subsidence slowed down and a regional disturbance accompanied by fault reactivation changed the depositional framework over large parts of the basin. During the Late Palaeocene the eastern part of the carbonate platform (Khalifa, Az Zahrah, Zaltan, Harash, and Kheir formations) continued to expand, whereas in the area west of the Marâdah trough block faulting gave rise to abrupt facies changes (Gumati and Kanis, 1985). During the Lower Eocene (Ypresian), large quantities of evaporites were deposited (Gir Formation; Hon evaporites), whereas throughout the Middle and Late Eocene carbonate deposition again prevailed (Gialo and Augila formations).

In the eastern Sirt Basin a significant change in
Fig. 3. Structural and stratigraphic cross section through the central part of the Sirt Basin based partly on well data and partly on seismic data (after Massa and Delort, 1984). See Fig. 2 for the location.
depositional environment occurred at the onset of the Oligocene which is related to uplift to the south, resulting in a northward retreat of the sea (Benfield and Wright, 1980). The sedimentary sequence changed to continental sands and clays (Arida and Diba formations) in the southern Sirt Basin while in the rest of the Sirt Basin marine deposition continued. A major regression occurred in post-Middle Miocene times due to a significant and rapid fall in the level of the Mediterranean (Barr and Walker, 1973). This marks the onset of continental deposition over large parts of the basin (Maradah Formation). In the area of the Hūn graben volcanic activity resumed during the Late Miocene.

QUANTITATIVE SUBSIDENCE ANALYSIS OF WELL DATA

Methods for subsidence analysis

Geohistory analysis is defined by Van Hinte (1978) as the use of quantitative stratigraphic techniques to unravel and portray geologic history. This technique uses microstratigraphical data concerning depositional environments expressed in terms of water depth and geologic ages in terms of million years instead of conventional terms. The burial plot is obtained by linearly interpolating between the depth at the present time for the first layer and the time at which this layer was on the surface. Ignoring compaction effects, all layers move up at the same rate as the first layer. The geohistory or basement subsidence curve results from plotting sediment accumulation in time and adding the palaeo-waterdepth at each time of deposition. This procedure continues until all layers are moved to the surface. However, to obtain a valid subsidence curve a correction for sediment compaction during deposition is necessary. To restore the sediment layer to its original thickness the effects of compaction of the section through time have to be removed. Van Hinte (1978) showed that the thickness of a unit at time of deposition and any time thereafter is related to the change in porosity of the sediment during burial. Since during burial the volume of the grains does not change, and the volume of pore space decreases significantly, the original thickness of a unit can be related to present-day thickness as follows:

\[ T_o = \frac{(1 - \Theta_o) T_n}{1 - \Theta_n} \]

where \( \Theta_o \) is the original porosity at the time of deposition and \( T_n \) and \( \Theta_n \) are the present-day thickness and porosity of the unit, respectively. In the absence of porosity data from wells, phenomenological models can be used to relate porosity to depth or load for various lithologies. Sclater and Christie (1980) proposed an exponential decrease of porosity with depth. Although the effects of compositional variations, over-pressure and diagenesis are not taken into account the model approximates the overall depth behaviour determined empirically from a variety of studies (e.g. Bond et al., 1983). The change of porosity during burial (\( \Theta_n \)) is given by

\[ \Theta_n = \Theta_o e^{-cZ} \]

where \( Z \) is the depth in metres, and \( \Theta_o \) the porosity when originally deposited (for \( Z = 0 \)). The constant \( c \) is a characteristic depth constant for each lithology. In these approaches, all changes of porosity are attributed to compaction thus ignoring the effects of early cementation. Once compaction has been removed from the original plot of sediment accumulation and the palaeo-bathymetric estimates for each stratigraphic unit are added, a curve representing the true basement subsidence results. This total subsidence curve includes the effects of sea-level changes, subsidence caused by sediment loading and tectonic subsidence. Tectonic subsidence is the undistorted basin subsidence that would have occurred in the absence of sedimentation and is therefore more directly related to the underlying origin of basin subsidence. Removing sediment loading and replacement of sediment infill by water is achieved by assuming Airy isostatic rebound of the basement. Examples of this technique of backstripping as discussed by Steckler and Watts (1978) can be found in Sclater and Christie (1980) and Bond and Kominz (1984). The calculation of tectonic subsidence in a unit of a stratigraphic section is given by

\[ Z_i = S' \left( \frac{\rho_a - \rho_i}{\rho_a - \rho_w} \right) + Wd_i - \Delta SL_i \left( \frac{\rho_a}{\rho_a - \rho_w} \right) \]

where \( Z_i \) is the tectonic subsidence relative to sea level, \( S' \) is the total sediment thickness under the top of unit \( i \) corrected for compaction, \( \rho_a \) is the density of the asthenosphere, \( \rho_i \) is the density of the sediment column, \( \rho_w \) is the density of water, \( Wd_i \) is the average water depth in which unit \( i \) was deposited, and \( \Delta SL_i \) is the sea-level change (rise is positive; fall is negative). In this study, water-loaded tectonic subsidence curves were calculated adopting the time scale of Harland et al. (1982).

Lithological effects, in particular compaction and extreme densities of evaporites, have been corrected for, using exponential decrease of porosity with depth (Bond et al., 1983). Each stratigraphic unit between two chrono-stratigraphic horizons has been assigned a
Table 2. Maximum and minimum compaction parameters for different lithologies used in the backstripping procedure (after: Bond and Kominz, 1984). $\Phi_1$ and $\Phi_0$ are the surface porosities (given as a fraction), $C_1$ and $C_0$ are the characteristic depth constants (in km$^{-1}$) for the deep and shallow porosity-depth relation respectively, $Z_p$ is the depth (in km) at which the deeper porosity-depth relation takes over from the shallow one, and $P_{gr}$ is the grain density (in g/cm$^3$) of the corresponding lithology.

<table>
<thead>
<tr>
<th>Lithology</th>
<th>$\Phi$</th>
<th>$C_1$</th>
<th>$C_0$</th>
<th>$Z_p$</th>
<th>$P_{gr}$</th>
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<tbody>
<tr>
<td><strong>Maximum</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
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<td>0.216</td>
<td>0.4</td>
<td>0.51</td>
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<tr>
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<td>0.375</td>
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<td>0.5</td>
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<tr>
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<td>0.475</td>
<td>0.7</td>
<td>1.1</td>
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<tr>
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<td>0.442</td>
<td>0.78</td>
<td>1.33</td>
<td>0.5</td>
</tr>
<tr>
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<td>0.10</td>
<td>0.0</td>
<td>0.10</td>
<td>0.0</td>
</tr>
<tr>
<td>Anhydrite</td>
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<td>0.10</td>
<td>0.0</td>
<td>0.10</td>
<td>0.0</td>
</tr>
<tr>
<td><strong>Minimum</strong></td>
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<td></td>
<td></td>
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<td>0.2</td>
<td>0.48</td>
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<tr>
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</tr>
<tr>
<td>Anhydrite</td>
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<td>0.10</td>
<td>0.0</td>
<td>0.10</td>
<td>0.0</td>
</tr>
</tbody>
</table>

sand, silt, shale, carbonate, and anhydrite percentage with each lithology responding according to its own compaction scheme. Porosity-depth relations were defined over two intervals, therefore two exponential expressions were necessary to describe the delithification process and a depth at which the deeper relation takes over from the shallow relation (see Bond and Kominz, 1984: p. 172):

$$
\Phi = \Phi_0 e^{-C_0z}, \quad 0 < z < Z_p
$$

$$
\Phi = \Phi_1 e^{-C_1z}, \quad Z_p < z < \infty
$$

where $z$ denotes the depth (in km), $\Phi_0$ and $\Phi_1$ denote the surface porosity (as a fraction), $c$ is the characteristic depth constant (in km$^{-1}$), and $Z_p$ denotes the depth at which the deeper porosity-depth relation takes over from the shallow relation. Minimum and maximum limits of compaction effects have been tested using the parameters listed in Table 2. These represent the range of values commonly observed in quantitative subsidence analysis (Bond and Kominz, 1984). The results of incorporating compaction effects showed that the uncertainties in basement and tectonic subsidence, related to the range in compaction parameters, are in the order of several tens of metres. However; drastic accelerations in basement and tectonic subsidence were observed related to variations in compaction behaviour; nevertheless, phases of rapid uplift cannot be explained as a result of compaction effects.

Palaeobathymetric data for the Late Cretaceous strata are taken from Megerisi and Mamgain (1980) and Eliaioubi and Powell (1980) who based their analyses on palaeo-environmental studies of planktonic and bentonic foraminifers. Using the Ingle (1980) palaeobathymetric scale these data were transferred into estimates of depositional depth. The accuracy of the palaeo-waterdepth estimates obtained according to this method is within 50 m for shallow-water deposits, but increases to 100–500 m for deeper water deposits.

Effects of long-term eustatic sea-level changes (after Kominz, 1984; see Fig. 4) on the inferred tectonic subsidence were investigated. Figure 5 shows the results of the analysis of the tectonic subsidence of well G1-47 applying the long-term sea-level curves of Kominz (1984). Although the Cenozoic fall in long-term sea level tends to decrease the overall tectonic subsidence rate, it appears to have little effect on the inferred subsidence pattern. In view of this, no corrections for long-term eustatic sea-level changes were applied to the other well analyses since they only affect the magnitude of the tectonic subsidence and have only a minor influence on the shape of the subsidence curves.

Airy isostasy was assumed, ignoring the effect of finite strength of the lithosphere on its response to sediment loading. This assumption, however, does not significantly affect the shape of the subsidence curves (Watts et al., 1982), particularly during the rift stage, when the lithosphere is mechanically weak (e.g. Hegarty et al., 1988). Similarly, we have ignored the reduced basement cooling resulting from the blanketing effect of
sediments (Turcotte and Ahern, 1977; Lucazeau and Le Douaran, 1985). Detailed knowledge of the thermal structure of the lithosphere throughout the evolution of the basin would be required to incorporate this effect into our analysis. However, because of the long time scale on which lithospheric cooling operates, a correction of the blanketing effect will not significantly alter the shape of the subsidence curves (Lucazeau and Le Douaran, 1985).

**Tectonic subsidence history of the Sirt Basin derived from well data**

Figure 6 shows curves for total basement subsidence for wells located on six different basement blocks. Overall subsidence of the basin began during the Albian and Cenomanian with relatively moderate subsidence rates (up to 7.5 cm/1000 yrs). During the Early Maastrichtian to Palaeocene subsidence accelerated to rates of 10 cm/1000 yrs. During the Eocene subsidence slowed down and continued at low rates until the Early Miocene.

Water-loaded tectonic subsidence curves constructed from well data are shown in Fig. 7. For each of these tectonic subsidence curves, plots of tectonic subsidence rates (in cm/1000 yrs) as a function of age (for 2 Ma intervals) have been constructed and grouped in accordance with the structural domains (Fig. 8). Inspection of Fig. 8 shows a fairly consistent subsidence and uplift history for the investigated basement blocks, regardless of whether these correspond to troughs or intrabasinal platforms.

The subsidence history of the West Zallah trough, which is based on wells C1-78 and H1A-57, is characterized by low subsidence during the Late Cretaceous. A subsidence peak marks the end of the Campanian, and a strongly undulating subsidence pattern is evident during the Palaeocene and Eocene, with a minor uplift occurring during the Ypresian between 54 Ma and 51 Ma. The Mid-Zallah trough shows a largely similar subsidence history to the West Zallah trough. Subsidence for this basement block was terminated during the Late Eocene and Oligocene. The East Zallah trough (based on wells B1-57, X1-11, P1-32, GG1-11
and J1-57) is characterized by low subsidence rates during the Cretaceous, and a subsidence peak during the Late Maastrichtian and Early Palaeocene (Danian). Small uplifts occurred at 61 and 53 Ma. Well J1-57 shows an anomalous pattern during the period 69–64 Ma. The subsidence history of the ‘Az Zahrah platform reveals a general trend towards uplift during the Early Tertiary with three minor uplift pulses at 61, 53 and 41 Ma. For the Marādah trough, wells F1-13, A1-94 and Y1-13 show low subsidence rates during the Late Cretaceous and increased subsidence at the end of the Maastrichtian. During the Palaeocene low subsidence rates were succeeded by small uplift pulses marking the end of the Eocene and the beginning of a period of almost zero subsidence. Due to the lack of extensive well data for the Zaltan platform, its subsidence history cannot be clearly ascertained. Wells 111-6 and FF1-6 show increasing subsidence during the Late Cretaceous and Palaeocene. A sudden uplift pulse occurs during the Early Eocene (55–53 Ma) and is succeeded by a pronounced subsidence peak during the period 47–45 Ma.
From the analysis of well data it is evident that all investigated basement blocks in the Sirt Basin show low subsidence rates during the Cretaceous. The end of the Cretaceous marks the beginning of a period of differential block movement which continued throughout the Palaeocene. A minor uplift pulse during the Early Eocene (Ypresian) followed by a subsidence peak during the Middle Eocene is recorded in all basement blocks. During the Late Eocene and Oligocene subsidence dies out and reaches very low rates.

**Fig. 7.** Water-loaded tectonic subsidence curves for wells covering six major basement blocks in the Sirt Basin.

**INTRAPLATE STRESSES AND THE POST-RIFT EVOLUTION OF THE SIRT BASIN**

**Tectonic subsidence history of investigated basement blocks**

Subsidence rate curves demonstrate that the tectonic and subsidence history of the Sirt Basin can be characterized by four distinct tectonic phases. A period of relatively low tectonic subsidence characterizes the Ceno-
Fig. 8. Tectonic subsidence rate curves for major basement blocks in the central part of the Sirt Basin.

Cretaceous to Campanian (phase 1). During the period from (Late Campanian) Early Maastrichtian to Palaeocene rapid subsidence indicates a first phase of extension, corresponding to the initiation of the Sirt Basin (phase 2). The East Zallah trough is characterized by a Late Maastrichtian to Early Palaeocene major subsidence period at which time it developed into a local depocentre, trapping large quantities of shale. During the late Danian and Early Thanetian (61–54 Ma) the East Zallah trough was uplifted but subsidence prevailed again toward the end of the Palaeocene. The area east of the Zallah trough is characterized by continuous subsidence during the Late Cretaceous and Palaeocene. The Az Zahrah platform shows rapid subsidence during the Late Maastrichtian and Early Palaeocene (68–63 Ma) and during the Late Palaeocene (57–54 Ma). The Maradh trough is characterized by high subsidence rates during Late Maastrichtian and slow subsidence at rates of 1 cm/1000 yrs during the Palaeocene. The Zaltn platform is characterized by continuous subsidence of 5 cm/1000 yrs during the Late Maastrichtian and Palaeocene. The differences in both magnitude and direction (e.g. uplift or subsidence) of movement of all the basement blocks with respect to each other is a clear indication of syn-sedimentary normal faulting typical during rifting events. Following this rifting phase is a period characterized by decelerating subsidence rates during the Eocene to Early Miocene. During the Eocene deviations from this subsidence pattern can be observed affecting the investigated basement blocks in the Sirt Basin (phase 3). The subsidence rates decelerate to very low levels during the Oligocene-Early Miocene (phase 4).
The overall pattern of tectonic subsidence observed during the Cenomanian-Early Miocene period is typical for a rift basin (McKenzie, 1978) where the period characterized by relatively low subsidence rates (Cenomanian-Campanian) represents the pre-rift period, the phase of rapid subsidence (Late Campanian-Palaeocene) represents the actual rifting phase during which the lithosphere isostatically responds to crustal thinning, followed by a post-rift phase with decelerating tectonic subsidence rates driven by thermal re-equilibration (Eocene-Early Miocene).

Estimates of crustal extension from simple stretching models

The amount of crustal extension can be obtained from a comparison of subsidence curves and estimates of the amount of crustal thinning from predictions of stretching models for basin formation (McKenzie, 1978; Royden and Keen, 1980). However, since lithospheric attenuation during rifting is also achieved by thermal upward displacement of the lithosphere-asthenosphere boundary (Ziegler, 1992), care should be taken in directly relating stretching factors to the magnitude of the thermal anomaly derived from quantitative subsidence analysis.

Stratigraphic and synthetic subsidence curves were modelled for individual well data using two-layer stretching models with an initial lithospheric thickness of 125 km and an initial crustal thickness of 31.3 km. In our model, the lithosphere was made up of blocks 10 km wide, each block characterized by a variable stretching factor. Thus a set of synthetic subsidence curves were generated modelling the basin evolution in time for various amounts of crustal and subcrustal extension. The thermal structure of the lithosphere was represented by a simple linear steady state geotherm. The temperature distribution within each of the extended blocks was taken as a function of the amount of stretching, the duration of rifting, the amount of stretching in neighbouring blocks, and the heat production (Kooi and Cloetinmph, 1989). Analytical solutions for thermal calculations during one instantaneous stretching event are given by Royden and Keen (1980) for vertical heat conduction. For multiple stretching events or stretching events of finite duration, a finite difference algorithm was adopted to derive the vertical component of heat conduction (Verwer, 1977). Deviations from the stretching model resulting from lateral heat flow (Cochran, 1983) were not incorporated because of a lack of data. From the thermal state of the lithosphere and the amount of crustal and sub-

<table>
<thead>
<tr>
<th>Parameter</th>
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<tr>
<td>Initial lithospheric thickness</td>
<td>$L$</td>
<td>125 km</td>
</tr>
<tr>
<td>Initial crustal thickness</td>
<td>$C$</td>
<td>31.3 km</td>
</tr>
<tr>
<td>Coefficient of thermal expansion</td>
<td>$a$</td>
<td>$3.4 \times 10^{-5}\text{C}^{-1}$</td>
</tr>
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<td>Asthenospheric temperature</td>
<td>$T_a$</td>
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</tr>
<tr>
<td>Thermal diffusivity</td>
<td>$\kappa$</td>
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</tr>
<tr>
<td>Mantle density</td>
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</tr>
<tr>
<td>Crustal density</td>
<td>$\rho_c$</td>
<td>2.8 gcm$^{-3}$</td>
</tr>
<tr>
<td>Isotherm describing EET</td>
<td>$T_e$</td>
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<tr>
<td>Water density</td>
<td>$\rho_w$</td>
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<tr>
<td>Surface porosity of sediment infill</td>
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</tr>
<tr>
<td>Characteristic depth constant</td>
<td>$c$</td>
<td>0.55 km$^{-1}$</td>
</tr>
</tbody>
</table>

The synthetic tectonic subsidence was calculated assuming Airy isostasy. Subsequently, the (unloaded) tectonic subsided basin was filled with water and sediments to give the water-loaded tectonic and basement subsidence, respectively, assuming Airy isostasy. Compaction of sediments was incorporated by using a standard porosity-depth relation $\Phi(z)=0.55 e^{-0.55z}$ for all sediments in the basin. Table 3 shows the parameters used in the thermo-mechanical models.

A simple stretching model was used first with an instantaneous uniform stretching phase at 75 Ma in which the crustal ($\delta$) and subcrustal ($\beta$) stretching factors had the same value ranging from 1.1 to 1.4. (Fig. 9). The synthetic tectonic subsidence reasonably approximates the observed thermally driven post-rift tectonic subsidence of the basin, but fails to fit the Late Cretaceous-Palaeocene syn-rift subsidence. Therefore a model with a stretching phase of finite duration during the Upper Cretaceous and Palaeocene (75–55 Ma) was used. Finite stretching rates were approximated by a large number of instantaneous stretching events of small magnitude, each followed by a short period of cooling (Jarvis and McKenzie, 1980). Equal crustal ($\delta$) and subcrustal ($\beta$) stretching factors ranging from 1.1 to 1.3 were used. Different water-depth profiles were used for the margin and the central part of the rifted basin. The model results, shown in Fig. 10, successfully predict the overall characteristics of the long-term patterns of tectonic subsidence during the Late Cretaceous and Tertiary. An almost linear tectonic subsidence pattern resulting from thinning and faulting of the brittle upper crust during the rift stage is clearly seen in the observed tectonic subsidence curves and is predicted by the stretching model. The observed thermally driven post-rift subsidence is well approximated by the results of the simple stretching models. However, short-term
Fig. 9. Comparison of observed tectonic subsidence (solid curve) and predicted synthetic subsidence (dashed curves) of the lithosphere using a two-layered stretching model (McKenzie, 1978; Royden and Keen, 1980) with an instantaneous rifting event at 75 Ma. Other parameters used are listed in Table 3. Individual wells showing the tectonic subsidence history of the six investigated basement blocks have been plotted for comparison. The instantaneous rifting model fails to account for the observed syn-rift subsidence during the Late Cretaceous-Palaeocene, although the thermally driven post-rift subsidence is approximated well by the model.

deviations from the predicted subsidence as seen in the Eocene cannot be attributed to thermal processes associated with basin formation. As discussed previously, these short-term deviations in the subsidence record have frequently been interpreted in terms of eustatic changes in sea level (e.g. Greenlee et al., 1988) but can also be explained by short-term changes in tectonic stress. The latter possibility was explored, focusing on the Eocene subsidence record which shows the most prominent deviations in subsidence pattern. Rift basins, generally, are characterized by a narrow graben structure overlain by a wider post-rift basin developing in response to thermal subsidence and sediment loading (e.g. Kooi and Cloetingh, 1989). This strong differential loading makes rifted basins very sensitive to intraplate stress fluctuations.

By comparing the observed tectonic subsidence and the predicted subsidence from stretching models, \( \beta \)-estimates were derived for different basement blocks. These estimates of crustal extension are in good agreement with data presented by Gumati and Nairn (1991). If we consider their estimate of 75\% extension for the Sirt trough, which is the central graben of the Sirt Basin, an average extension of 40\% is found for the Sirt Basin as a whole on the basis of simple stretching models. Crustal stretching estimates for platform areas are smaller than for the basinal areas and increase towards the central part of the basin. Ziegler (1992), however, pointed out that extension-induced thermal and convective processes contribute to thinning of the subcrustal lithosphere, associated with magmatic underplating and injection of mantle derived melts into the crust (Morgan and Ramberg, 1987). Thus, lithospheric attenuation during rifting is not achieved purely by extension, but also involves upward displacement of the asthenosphere-lithosphere boundary. Phase transformations of crustal rocks may also contribute to the post-rift subsidence of basins. Therefore, care should be taken in
deriving crustal stretching factors from subsidence analysis of post-rift basins, and literally translating these β-estimates into conclusions bearing on the formation of rifted basins.

**Effects of late-stage compression on estimates of crustal thinning**

Simple stretching models (McKenzie, 1978; Royden and Keen, 1980) predict the subsidence history of a basin formed by extension. These models assume that the upper crust responds to a period of stretching by faulting and thinning, therefore another independent estimate of crustal extension can be obtained by measuring horizontal displacements along normal faults. Stretching estimates derived from horizontal displacement of normal faults resulted in an average crustal extension of 23% over the entire basin. This estimate is based on seismic and stratigraphic cross sections published by Conant and Goudarzi (1967), Goudarzi (1980), Goudarzi and Smith (1978), Gumati and Kanes (1985), and Gumati and Nairn (1991) as well as on our own data. As discussed previously, quantitative modelling of the observed tectonic subsidence using simple stretching models has yielded an estimated extension of about 40%. Therefore a discrepancy exists between stretching estimates obtained from synthetic models and stretching estimates derived from horizontal slip along upper crustal normal faults. This discrepancy is generally recognized in basin studies (Moretti and Pinet, 1987). For example, quantitative modelling of the observed subsidence for the central graben of the North Sea Rift basin by Barton and Wood (1984) yielded an extension of about 50–80 km, whereas Ziegler (1982) pointed out that the total amount of crustal extension, measured on high-angle normal faults, is probably only 20–25 km. Current backstripping techniques, correcting only for vertical loading of the lithosphere, tend
to overestimate the amount of extension. As pointed out by Kooi and Cloetingh (1989), stress-induced basin subsidence may explain at least part of the observed discrepancy since the incorporation of late-stage compression is equivalent to adopting an increased estimate of $\beta$. In situ stress measurements by Schäfer et al. (1980) have also shown a change in the stress regime from NE-SW tensional to NW-SE compressional stress in the Neogene, whereas Sestini (1984) identified a post-Eocene phase dominated by compressional events throughout the African margin. Gealey (1988) showed that first signs of compression date back to the Middle Eocene. According to Guiraud et al. (1992), intra-Eocene compressional events are evident from the sedimentary fill of most basins of the west and central African Rift System. They attribute these compressional events to a major stage in the collision between the African and European plates, known as the Pyrenean-Atlasic phase, which is directly correlated with a change in opening direction of the Central Atlantic. Evidence for stress-induced basin subsidence resulting from late-stage compression is also observed in the Late Cenozoic evolution of the North Sea basin (Kooi et al., 1991; Ziegler and Van Hoorn, 1989). Similarly, a number of rifted margins in the Mediterranean are to a large extent dominated by Late Neogene increase in subsidence that can be related to late-stage compression as a result of the Africa-Europe collision (e.g. the Gulf de Lions: see Burrus et al., 1987; Levant margin: see Tibor et al., 1992).

**DISCUSSION AND CONCLUSIONS**

Figure 11 summarizes the timing of tectonic events in the Sirt Basin in comparison with the changes in relative motion vectors between the African and European plates documented by Dewey et al. (1974), Savostin et al. (1986), and Ziegler (1988). Palaeostress data derived from Schäfer et al. (1980) are also included.

Burke and Dewey (1974) suggested that rifting
activity in the Sirt Basin started in the Early Cretaceous and was related to dextral shear along the Gibraltar-Maghrebian shear zone thus forming a diffuse trans-tensional plate boundary between Africa and the Italo-Dinaride block. Guiraud and Maurin (1991) suggested a Neocomian-Barremian onset of rifting resulting in the development of dominantly E–W trending half-grabens associated with N140E trending faults for the southern Sirt Basin. Bayoumi and Lofty (1989) found similar structures in the Abu Gharadig Basin located in the Western Desert of Egypt. This Early Cretaceous rifting event is related to progressive northward propagation of crustal separation between Africa and South America and to changes in rates of sea-floor spreading in the central Atlantic (Guiraud et al., 1992). A second stage of rifting, initiated at the end of the Early Aptian and lasting until Late Albian times, resulted in NW–SE trending rifts evolving in response to a NE–SW crustal extension (Guiraud et al., 1992).

Differential rates of sea-floor spreading in the central and South Atlantic and contemporaneous rapid opening of the equatorial Atlantic Ocean provide an explanation for the rejuvenation of crustal zones of weakness in Africa (Fairhead and Binks, 1991).

According to the present analysis, active syn-rift subsidence in the northern part of the Sirt Basin started in Late Campanian (± 75 Ma) and continued until Late Palaeocene. Savostin et al. (1986) and Le Pichon et al. (1988) documented a change in the relative motion vector of the African and European plates from left-lateral translation to convergence in the Campanian (± 80 Ma) which coincides with the initiation of rifting observed from the subsidence analysis.

The Eocene-Early Miocene tectonic history of the Sirt Basin shows decelerating tectonic subsidence rates typical for the post-rift evolution of rift basins driven by thermal contraction. Results of synthetic modelling of the tectonic subsidence history showed that simple
stretching models predict the initial syn-rift subsidence and the overall characteristics of the long-term patterns of tectonic subsidence. These models, however, fail to account for short-term deviations from the predicted subsidence pattern such as the sudden uplift in the Early Eocene and the sudden subsidence in the Late Eocene. Since these short-term deviations from the predicted subsidence pattern cannot be attributed to thermal processes, we interpret them as resulting from changes in intraplate stress levels. The well data we have used in our analysis cover mainly the northwestern margin of the Sirt Basin and do not extend into the central graben (the Sirt trough) of the basin. Therefore, when applying the model for flexural deflections of rift basins presented in Fig. 1, the observed uplift in Early Eocene could have resulted from an increase of the level of horizontal intraplate compression whereas the subsequent Late Eocene subsidence can be explained by a sudden stress release or intraplate extension yielding isostatic rebound of the lithosphere. Guiraud et al. (1992) discuss intra-Eocene compressional events, referred to as the Pyrenean-Atlasic phase, resulting from changes in the level of intraplate compression. They relate these phases to changes in the Atlantic flowline geometries, reflecting a change in opening direction of the Central Atlantic (Fairhead and Binks, 1991) and consequently the drift pattern of Africa.

This phase of basin uplift and subsidence abruptly terminated during the Late Eocene as a result of the change in the motion vector of Africa relative to Europe from compressional to right lateral movement (see Savostin et al., 1986). These authors attribute this change in motion vector to an abrupt increase in spreading rate between the African and North American plates. An Oligocene rift phase is evident for many East African basins (e.g. Tenere, Sudan, and Anza basins: Genik, 1992; McHargue et al., 1992). Although the geodynamic processes underlying this event are not clear, Guiraud et al. (1992) discuss the resumption of crustal stretching in terms of a renewed intraplate distensive stress regime. In the sedimentary fill of the Sirt Basin this stretching event is evidenced by a widespread unconformity (Benfield and Wright, 1980; Barr and Weegar, 1972) marking the onset of a period of low and decelerating subsidence rates. Termination of a phase of low subsidence rates and marine deposition in the Early Miocene coincides with the final change to northeasterward compressional motion of Africa relative to Europe (Savostin et al., 1986; Livermore and Smith, 1985) accompanied by a rotation of the palaeostress field to ENE–WSW compressional during the Early-Middle Miocene and N–S during the

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<th>INTRAPLATE STRESS FIELD</th>
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<td>(After: Harland et al. 1982)</td>
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<td>COMPRESSION (kbars)</td>
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<tr>
<td>MOCONIAN</td>
<td></td>
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<tr>
<td>SIERRAVALLIAN</td>
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<tr>
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<td>EARLY LANGHIAN</td>
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<td>CAMPAICANIAN</td>
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Fig. 12. Palaeostress curve inferred from basin stratigraphy incorporating the results from subsidence analysis.

Late Miocene (Le Pichon et al., 1988). Palaeostress analysis by Schäfer et al. (1980) showed a rotation of the stress field from NE–SW extensional to a NW–SE compressional regime during the Early Miocene. This event coincides with the opening of the southern Hün graben and contemporaneous magmatic activity during the Late Miocene and Pleistocene (Ade-Hall et al., 1974; Wilson and Guiraud, 1992).

Figure 12 shows the palaeostress field inferred from the quantitative subsidence analysis in conjunction with synthetic modelling of basin formation using simple stretching models (McKenzie, 1978; Royden and Keen, 1980). The long-term trend of the palaeostress pattern is that of a change from overall tension during the Cretaceous and Early Tertiary to a stress regime dominated by compression during the Late Tertiary. Superimposed on this trend are more abrupt changes. The palaeostress
curve suggests that termination of the rifting and differential faulting phase is associated with major relaxation of tensile stresses followed by renewed accumulation of tension.

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