Lütting, S.

LATE CRETACEOUS - EARLY TERTIARY SEQUENCE STRATIGRAPHY, PALEOECOLOGY AND GEODYNAMICS OF EASTERN SINAI, EGYPT
The "Berichte aus dem Fachbereich Geowissenschaften" are produced at irregular intervals by the Department of Geosciences, Bremen University. They serve for the publication of experimental works, Ph.D.-theses and scientific contributions made by members of the department.

Reports can be ordered from:

Gisela Boelen
Sonderforschungsbereich 261
Universität Bremen
Postfach 330 440
D 28334 BREMEN
Phone: (49) 421 218-4124
Fax: (49) 421 218-3116

Citation:
Lüning, S.
Late Cretaceous - Early Tertiary sequence stratigraphy, paleoecology and geodynamics of Eastern Sinai, Egypt.
Bremen, 1997

ISSN 0931-0800
Late Cretaceous - Early Tertiary
sequence stratigraphy, paleoecology and
geodynamics of Eastern Sinai, Egypt

Dissertation
for the doctorate degree
of the Department of Geosciences
at the University of Bremen

submitted by
Sebastian Lüning
1997
Tag des Kolloquiums:
29. 9. 1997

Gutachter:
Prof. Dr. J. Kuss
Prof. Dr. M. Olesch

Prüfer:
Prof. Dr. O. Brockamp
Prof. Dr. H. D. Schulz
I shall now come to speak of Egypt, for it holds a
great many marvels and more than any other
country offers monuments almost beyond
description.

*Herodotus*
I am extremely happy about the effective and smooth collaboration with Dr. Akmal M. Marzouk (Tanta University, Egypt). He provided all data and illustrations of calcareous nanofossils, and by that he expanded enormously the dimensions of the project. The results of his quick and comprehensive investigations allowed many interpretations which could not have been accomplished from foraminifera data alone.

I also would like to thank Dr. A. Mohsen Morsi for his detailed ostracod investigations, his logistical help in Cairo and for various discussions. From Robert Speijer (formerly Goeteborg University, now Bremen University) I received valuable help with the identification and paleobathymetric interpretation of benthonic foraminifera. His critical comments on an earlier version of chapter 6 improved the paper significantly. For detailed remarks on chapter 7, two anonymous reviewers are acknowledged.

Special thanks go to Prof. A. M. Bassiouni who always had to work hard before the various expeditions were ready to start to the field, and to Dr. Peter Luger whose discussions and help with the foraminiferal biostratigraphy were most valuable for me. For various kinds of help and encouragement I am indebted to Dr. Kai-Uwe Gräfe, Ralf Gietl, Dr. Frank Kulbrok, Dr. Martina Bachmann, Kristin Schnack, Christian Scheibner, Dr. Joachim Jacobs, Tanja Hildebrand-Habel, Dr. André Freiwald (all University of Bremen), Dr. Mohammed El Beshtawy (Benha University, Egypt), Dr. Amin Strougo (Ain Shams University, Cairo), Prof. Mahmoud Kora (Mansoura University, Egypt), Prof. Noel Vandenberghe (KU Leuven), Enrique Bernades Rodriguez (Univ. Oviedo, Spain), and M. Wiese (FU Berlin).

During the course of the study I received considerable technical help from Markus Brinkmann, Dr. Hartmut Mai, Ralph Kunow, Jürgen Beyer and Katrin Virkus whom I thank for their hours and days devoted to photography and sample preparation. The English language in this contribution was significantly improved by Mrs. Erna Friedel (Bremen), Prof. W. R. Fitches (Aberystwyth) and Dr. Linda Franklin (Helgoland) who read parts of the manuscript.

Funding for the project was provided by the Deutsche Forschungsgemeinschaft (DFG grant Ku 642/8-1). The DFG also provided funding for the participation at two international congresses, while further financial support came from the University of Bremen (FNK) and the ‘Graduiertenkolleg’.

Above all, I thank my parents for their extraordinary support, both financially and morally. During all my university education they have been a constant source of encouragement. Finally, a large bouquet of flowers goes to my wife Renata. I am grateful for her unselfish technical assistance in various ways, her continuous moral support and patience during the busy episodes of the project.
Abstract

Twentyseven sections with a total length of about 2400 m were measured in eastern Sinai, covering intervals between the uppermost Cenomanian and middle Eocene. A total of 612 marly samples and 80 handspecimens have been collected. More than half of the sections are located on the tectonically stable shelf in central east Sinai while the rest are situated on the tectonically unstable shelf in northern Sinai, with most of them at the Gebel Areif El Naqa anticline. Field recordings and laboratory data from microfossils and microfacies form the basis for five papers dealing with sequence stratigraphy, paleoecology, biostratigraphy and geodynamics within a high resolution biostratigraphic frame mainly provided by planktonic foraminifera and calcareous nannofossils. Detailed abstracts of the publications can be found at the beginning of chapters 5-9.

The Turonian to Paleocen sedimentary succession of the tectonically quiet area (central east Sinai) was formed under terrestrial to hemipelagic conditions and has been influenced by a number of characteristic sea-level changes. While the Late Cretaceous paleobathymetric variations are documented by distinct facies breaks or hiatuses, the Paleocene sea-level history can be reconstructed mainly on the basis of the foraminiferal planktonic-benthonic ratio. Correlation with other sea level reconstructions from the region points to a more or less synchronous regional sea level development for the whole Turonian-Paleocene interval. In terms of sequence stratigraphy, six major 2nd-3rd order sequence boundaries have been reconstructed for the Turonian to Maastrichtian interval (chapter 5), while for the uppermost Maastrichtian to lower Eocene interval, ten 3rd order sequence boundaries have been interpreted (chapter 7). Comparisons of the studied Late Cretaceous sequences with the "eustatic" model involves uncertainties; nevertheless, some of the sea-level fluctuations recorded in Sinai may be correlated to worldwide eustatic sea level changes. During the late Maastrichtian Abathomphalus mayaroensis Zone, abundances of calcareous nannofossils and foraminifera were predominantly controlled by changes in paleoproductivity, as semiquantitative investigations have shown (chapter 6). Two microfossil assemblages are interpreted to reflect low and high surface water productivity conditions. The microfossil distribution patterns of these late Maastrichtian hemipelagites are strongly dominated by the southern Tethyan upwelling system, which was reported to have been active from the Santonian to the Late Maastrichtian. Termination of upwelling just before the K/T boundary also provides a good explanation for the change towards a paleobathymetric control on foraminiferal distribution as observed in the Paleocene succession. The early and mid Paleocene sea level curve from central east Sinai is in good correspondence with the sea-level history in Egypt, Tunisia, the European Basins, Texas, and the "eustatic" sequence chart, suggesting a eustatic control on deposition during this period in the study area. Furthermore, the Paleocene hemipelagites of Sinai were apparently deposited in a paleobathymetric and paleoceanographic setting suitable for paleodepth reconstructions on the basis of the foraminiferal planktonic-benthonic ratio.

The comparison of the biozonal distribution of Paleocene planktonic foraminifera and calcareous nannofossils in eastern Sinai yields a consistent regional correlation pattern (chapter 8). Nevertheless, when compared to the schemes from other regions, a great variability in relative timing of the foraminiferal and nannofossil bio-events can be observed throughout the different studies. A number of potential reasons for the shifts observed exist. The study suggests that interbasinal and sometimes even intrabasinal correlations using nannofossil and/or foraminiferal biostratigraphic data, must take similar (diachronous) variabilities into account.

The detailed sequence stratigraphic results from the tectonically stable area in central east Sinai (chapters 5, 7), the differentiation of control mechanisms on sedimentation (chapter 6), and biostratigraphic correlations (chapter 8) provided an important basis for the reconstruction of the syndepositional deformation at the Gebel Areif El Naqa anticline in north eastern Sinai (chapter 9). This domal anticline represents a key area of the 'Syrian Arc' which is considered an intraplate orogen formed by inversion of an older halfgabnen system as a consequence of the collision of the African and Eurasian Plates since the Late Cretaceous. The pre- and syndeformational late Albian to early Eocene depositional history at Areif El Naqa has been reconstructed in terms of sequence stratigraphy on the basis of detailed sedimentologic, biostratigraphic and paleoecologic investigations in ten sections as well as literature data. Three significant uplift phases during the studied period have been determined based on lateral facies and thickness changes, local development of pronounced hiatuses and comparison with the sequence stratigraphic development in the tectonically quiet regions further south. The uplift history at Areif El Naqa has been compared with the tectonic development in other parts of the Syrian Arc and in general seems to reflect major movements which occurred throughout the anticlines of the foldbelt.
Zusammenfassung

Im östlichen Sinai wurden 27 Profile mit einer Gesamtlänge von knapp 2400 m aufgenommen, die das stratigraphische Intervall vom oberen Cenoman bis zum mittleren Eozän abdecken. Dabei wurden 612 Schlämmproben und 80 Handstücke bearbeitet. Mehr als die Hälfte der Profile befinden sich auf dem tektonisch stabilen Schelf im zentralen Ost-Sinai, während die restlichen auf dem instabilen Schelf im nördlichen Ost-Sinai liegen. Das Untersuchungsmaterial bildet die Grundlage für fünf Publikationen, die sich mit der regionalen Sequenzstratigraphie, Paläoökologie und Geodynamik befassen, wobei planktonische Foraminiferen und kalkiges Nannoplankton den biostratigraphischen Rahmen definieren. Die Kurzfassungen der Artikel sind den Kapiteln 5-9 vorangestellt; eine Zusammenfassung der regionalen geodynamischen Entwicklung befindet sich in Kapitel 2.


Die paleozoänene Studie befaßt sich zudem mit der Anwendung des sequenzstratigraphischen Konzepts im hemipelagischen Bereich.


Contents

Preface

Abstract

1. Introduction
  1.1. Topography and geomorphology of Sinai ........................................ 1
  1.2. History of geological research on Sinai ........................................ 1
  1.3. Fieldwork ................................................................. 2

2. Geodynamic development of the Sinai Peninsula and
neighbouring regions during the Mesozoic and Cenozoic
  2.1. Summary ................................................................. 5
  2.2. Transcontinental shear system ............................................ 5
  2.3. Eastern Mediterranean Basin ............................................ 5
    2.3.1. Transpressional basin in central north Egypt .................... 8
    2.3.2. Rifting of the Turkish-Apulian terrane ......................... 8
    2.3.3. Syrian-Arc intraplate folding .................................. 12
    2.3.4. E-W strike slip of the Themed Fault .......................... 17
  2.4. Opening of the Red Sea and related processes ..................... 18
    2.4.1. Opening of the Red Sea, Gulf of Suez and Gulf of Aqaba .... 18
    2.4.2. Movements and structures at the Dead Sea Fault ............. 20
    2.4.3. Movements along the Sinai Hinge Belt and late Tertiary faults 21
    2.4.4. Neogene folds in the Red Sea fault systems ................ 21
    2.4.5. Intrusion of dykes ............................................. 21
  2.5. Importance of the tectonic processes for
      Cretaceous-Paleogene basin analysis in Sinai ................... 22

3. Regional sedimentary setting
  3.1. Stratigraphic overview ............................................. 23
  3.2. Hydrocarbon provinces in Sinai ..................................... 23
    3.2.1. Gulf of Suez Rift Basin .................................... 23
    3.2.2. Mesozoic Basin in the Eastern Mediterranean ............... 25

4. General concepts
  4.1. Sequence stratigraphic concept ..................................... 27
  4.2. Paleoecologic interpretations based on foraminifera and calcareous nannofossils 32

5. Sequence stratigraphy of the Late
Cretaceous of central east Sinai, Egypt
(S. Lüning, A. M. Marzouk, A. M. Morsi, J. Kuss) ......................... 33
  5.1. Introduction ......................................................... 33
  5.2. Regional setting .................................................... 33
  5.3. Materials and methods ............................................. 35
  5.4. Biostratigraphy ...................................................... 35
  5.5. Facies zones .......................................................... 41
  5.6. Sequence stratigraphy .............................................. 58
  5.7. Comparison with sea-level reconstructions from neighbouring regions 66
  5.8. Conclusions ........................................................... 68

6. Late Maastrichtian litho- and ecocycles
from the hemipelagic of eastern Sinai, Egypt
(S. Lüning, A. M. Marzouk, J. Kuss) .................................. 70
  6.1. Introduction .......................................................... 70
  6.2. Regional setting .................................................... 71
  6.3. Materials and methods ............................................. 71
  6.4. Biostratigraphy ...................................................... 72
  6.5. Paleoecologic parameters studied .................................. 72
6.6. Results ........................................................................................................ 75
6.7. Discussion .................................................................................................... 86
6.8. Conclusions ................................................................................................ 88

7. The Paleocene of central east Sinai, Egypt:
   ‘Sequence Stratigraphy’ in Monotonous Hemipelagites
   (S. Luning, A. M. Marzouk, J. Kuss) ................................................................. 90
   7.1. Introduction ............................................................................................... 90
   7.2. Regional setting ........................................................................................ 91
   7.3. Methods ..................................................................................................... 91
   7.4. Paleodepth indicators .............................................................................. 94
   7.5. Sequence stratigraphic concept ............................................................... 94
   7.6. Paleocene sequences in central Sinai ..................................................... 95
   7.7. Control mechanisms: Eustasy vs. tectonics ............................................ 105
   7.8. Conclusions ............................................................................................. 108

8. Comparative biostratigraphy of calcareous
   nannofossils and planktonic foraminifera in
   the Paleocene - early Eocene of eastern Sinai, Egypt
   (A.M. Marzouk, S. Luning) .............................................................................. 109
   8.1. Introduction ............................................................................................... 109
   8.2. Regional setting ........................................................................................ 109
   8.3. Methods ..................................................................................................... 110
   8.4. Biostratigraphy ........................................................................................ 110
       8.4.1. Planktonic Foraminifera .................................................................. 111
       8.4.2. Calcareous Nannofossils .................................................................. 129
       8.4.3. Biozonal correlation of planktonic foraminifera and calcareous nannofossils .................................................................................... 130
   8.5. Conclusions ............................................................................................. 131

   pre- to syndeformational deposition at the Areif El Naqa anticline
   (northern Sinai, Egypt).
   (S. Luning, J. Kuss, M. Bachmann, A. M. Marzouk, A. M. Morsi) ................. 133
   9.1. Introduction ............................................................................................... 133
   9.2. Geodynamic setting ................................................................................ 135
   9.3. Material and methods ............................................................................ 135
   9.4. Biostratigraphy ....................................................................................... 136
   9.5. Facies analysis and paleobathymetry ..................................................... 137
   9.6. Depositional history ................................................................................ 137
   9.7. Comparison with the depositional history on the stable shelf
       and reconstruction of the uplift history at Gebel Areif El Naqa ................ 159
   9.8. Comparison with the deformatonal history in other parts of the Syrian Arc .......................................................... 162
   9.9. Conclusions ............................................................................................. 165

REFERENCES .............................................................................................. 166

APPENDIX
   A Compilation of references for the Cenomanian-Eocene of Sinai (sorted by area) ...... 183
   B Summary of biozonal schemes of plankt. foraminifera and calcareous nannofossils ...... 187
   C Accumulation rates (late Maastrichtian-Paleocene) ........................................ 189
   D Sections ........................................................................................................ 191
1. Introduction

1.1. Topography and geomorphology of Sinai

Sinai has an almost triangular shape and is separated from the Egyptian mainland in the west by the Gulf of Suez and the Suez Canal, while the eastern boundary is formed by the Gulf of Aqaba and the Israeli State border (Fig. 1-1). The Mediterranean and the Red Sea mark the northern and southern boundaries, respectively. In a more general sense, the geographic term ‘Sinai Peninsula’ also includes the Israeli Negev Desert which is very similar to the Egyptian part in terms of geology, topography and ecology (Jahn & Jahn, 1994). The Sinai Peninsula occupies an area of more than 60,000 km^2 with a northern east-west extension of about 200 km and a north-south extension of approximately 380 km (Jahn & Jahn, 1994). Sinai forms an important landbridge between Africa and Asia and officially belongs to the Asian continent (Statistisches Bundesamt, 1993).

The arid peninsula can be subdivided into several geographic-geologic provinces. In the north, the highly saline Bardawil lagoon is protected from the open Mediterranean Sea by a barrier island. The northern coastal stretch of Sinai is dominated by a 20-50 km wide dune belt formed by the prevailing northern and northwestern winds (Jahn & Jahn, 1994). While northern Sinai is characterized by domal anticlinal belts, most of central Sinai is occupied by the Tih Plateau which reaches heights up to 1000 m. Large parts of southern Sinai are dominated by steep mountains with heights up to 2500 m made from colourful crystalline basement. Although wide areas of Sinai are occupied by desert, Sinai plays a significant role in Egyptian economy because of rich oil and gas reserves in the Gulf of Suez including west Sinai, and of extensive tourism in south and south-east Sinai which is associated with the well developed coral reefs in the Gulf of Aqaba and the northern Red Sea.

1.2. History of geological research on Sinai

ca. 2000 B.C. During a three months expedition to Sinai, led by Haroeris, large amounts of turquoise were discovered and mined (EGPC, 1986).

1822-1831 The german zoologist E. Rüppell travels to Sinai and conducts experiments of copper melting in Bir Nasib (Rothenberg, 1979).

1886 Discovery of oil at the western edge of the Gulf of Suez (Gemsa Field) which marked the initiation of production in the hydrocarbon rich Gulf of Suez basin (EGPC, 1986).


1921 Publication of a comprehensive summary of the geology of Egypt (‘Handbuch der Regionalen Geologie, Ägypten’) by the German Max Blanckenhorn (Blanckenhorn, 1921) who has worked on the egyptian geology since 1886.


Early 1960s Seismic exploration in several regions of Northern Sinai incorporated in a seismic reconnaissance survey for the Mediterranean offshore region (El Ayouty 1990: 596).


1967-1982 Following the six days war in 1967, the Geological Survey of Israel carried out extensive geological investigations on Sinai during the period of Israeli occupation, with a special
emphasis on hydrocarbon exploration. 15 holes were drilled, including three in the offshore area (Bartov & Katz, 1978; Said, 1990a; Sestini 1995; Alsharhan & Salah, 1996). During this campaign, the Sadot gas field, located in the extreme northeast of Sinai (SW Rafah), was discovered. Today, this reservoir in Cenomanian dolomites and limestones is almost exhausted (El Ayouty 1990: 596). During the 1973 Yom Kippur war part of Sinai is recaptured by Egypt. Following a peace treaty in 1979, Sinai is stepwise returned to Egypt until 1982.

1976
Foundation of the Egyptian General Petroleum Corporation (EGPC).

1980s
A new round of petroleum exploration focuses on the Mediterranean coastal strip and on Northeast Sinai. Despite a few oil- and gas shows (Tineh, Port Fouad, Abu Zakin, Wakar, South Rafah, Mango wells), the campaign is unsuccessful (Sestini 1995; Alsharhan & Salah, 1996).

1.3. Fieldwork

During two field campaigns in 1995 and 1996, a total of 27 sections with a length of 2395 m covering intervals between the uppermost Cenomanian and middle Eocene were measured (Tab. 1-1). Special attention had been paid to lithological changes and sedimentary structures. Sampling was focussed on pelitic horizons with potential to contain foraminifera and calcareous nannofossils. Handspecimens were taken from selected intervals to support the macroscopic field data for the reconstruction of the sedimentary facies. A total of 612 marly samples and 80 handspecimens have been collected (Tab. 1-1). From eleven sections located on the tectonically unstable shelf in northern Sinai, eight sections have been measured in the relatively small area of the Gebel Areif El Naqa anticline (Fig. 1-1). The remaining 16 sections are situated on the tectonically stable shelf in central east Sinai and were taken at erosional scarps cut by wadis and at the margins of blocks that had been tilted in connection with the Gulf of Aqaba rifting. While some of the sections are located at or close to the road or track (B1/B2, C, L1, L2, M, N2/3, T1/2, W) others needed longer offroad passages (A1-8, D, F, G, H, K, N1, P, Q, R). In particular, access to the sections on the Tih Plateau (Q, R) is highly complicated because the track is unmarked and is locally interrupted by wadis with coarse bedload. North of Themed, in the vicinity of section F, an unearthed tank mine has been found on the wadi ground which suggests the presence of mine fields in this area so that it may be advisable to avoid extensive offroad travelling here.
Fig. 1-1. Location map of sections measured in eastern Sinai.
<table>
<thead>
<tr>
<th>Section No.</th>
<th>Locality</th>
<th>Stratigraphic range</th>
<th>Length of section (m)</th>
<th>Number of marly samples (washed)</th>
<th>Number of handspecimens (thin sections)</th>
<th>Year of fieldwork</th>
</tr>
</thead>
<tbody>
<tr>
<td>A1</td>
<td>Gebel Areif El Naqa</td>
<td>upper Turonian-?Eocene</td>
<td>280</td>
<td>37</td>
<td>19</td>
<td>1995+96</td>
</tr>
<tr>
<td>A2</td>
<td>Gebel Areif El Naqa</td>
<td>Maastrichtian-Paleocene</td>
<td>69</td>
<td>10</td>
<td>2</td>
<td>1995</td>
</tr>
<tr>
<td>A3</td>
<td>Gebel Areif El Naqa</td>
<td>lower Turonian-?Coniacian</td>
<td>79</td>
<td>9</td>
<td>3</td>
<td>1995</td>
</tr>
<tr>
<td>A4</td>
<td>Gebel Areif El Naqa</td>
<td>upper Paleocene-Eocene</td>
<td>104</td>
<td>10</td>
<td>0</td>
<td>1995</td>
</tr>
<tr>
<td>A5</td>
<td>G. Areif El Naqa</td>
<td>upper Maastrichtian-Eocene</td>
<td>55</td>
<td>26</td>
<td>3</td>
<td>1996</td>
</tr>
<tr>
<td>A6</td>
<td>G. Areif El Naqa</td>
<td>Santonian-Maastrichtian</td>
<td>70</td>
<td>25</td>
<td>0</td>
<td>1996</td>
</tr>
<tr>
<td>A7</td>
<td>G. Areif El Naqa</td>
<td>Maastrichtian-Eocene</td>
<td>120</td>
<td>26</td>
<td>0</td>
<td>1996</td>
</tr>
<tr>
<td>A8</td>
<td>G. Areif El Naqa</td>
<td>upper Maastrichtian-Eocene</td>
<td>50</td>
<td>24</td>
<td>0</td>
<td>1996</td>
</tr>
<tr>
<td>C</td>
<td>5 km N' Sheikh Attiya</td>
<td>?Cenomanian-?Coniacian</td>
<td>86</td>
<td>44</td>
<td>1</td>
<td>1995+96</td>
</tr>
<tr>
<td>D</td>
<td>20 km N' Sheikh Attiya</td>
<td>Coniacian-Santonian</td>
<td>33</td>
<td>5</td>
<td>0</td>
<td>1995</td>
</tr>
<tr>
<td>E</td>
<td>30 km E' Themed</td>
<td>middle Eocene</td>
<td>134</td>
<td>26</td>
<td>0</td>
<td>1995</td>
</tr>
<tr>
<td>F</td>
<td>10 km NNE' Themed</td>
<td>upper Maastrichtian-Eocene</td>
<td>120</td>
<td>51</td>
<td>1</td>
<td>1995</td>
</tr>
<tr>
<td>G</td>
<td>25 km SE' Themed</td>
<td>?Turonian-?Coniacian</td>
<td>42</td>
<td>4</td>
<td>9</td>
<td>1995</td>
</tr>
<tr>
<td>H</td>
<td>30 km SE' Themed</td>
<td>?Turonian</td>
<td>57</td>
<td>4</td>
<td>3</td>
<td>1995</td>
</tr>
<tr>
<td>K</td>
<td>10 km W' Themed</td>
<td>Paleocene-lower Eocene</td>
<td>44</td>
<td>20</td>
<td>0</td>
<td>1995</td>
</tr>
<tr>
<td>L1</td>
<td>Bir Hasana</td>
<td>Paleocene-lower Eocene</td>
<td>88</td>
<td>21</td>
<td>1</td>
<td>1995</td>
</tr>
<tr>
<td>L2</td>
<td>Bir Hasana</td>
<td>middle Eocene</td>
<td>92</td>
<td>33</td>
<td>3</td>
<td>1998</td>
</tr>
<tr>
<td>M</td>
<td>Taba</td>
<td>upper Turonian-Eocene</td>
<td>240</td>
<td>45</td>
<td>15</td>
<td>1995+96</td>
</tr>
<tr>
<td>N1</td>
<td>N'Themed</td>
<td>?Turonian</td>
<td>65</td>
<td>5</td>
<td>1</td>
<td>1996</td>
</tr>
<tr>
<td>N2/3</td>
<td>NW'Themed</td>
<td>?Coniacian-?Santonian</td>
<td>42</td>
<td>4</td>
<td>0</td>
<td>1996</td>
</tr>
<tr>
<td>P</td>
<td>W. Gureis, S' Themed</td>
<td>upper Maastrichtian-Eocene</td>
<td>105</td>
<td>35</td>
<td>0</td>
<td>1996</td>
</tr>
<tr>
<td>Q</td>
<td>Egma Plateau</td>
<td>Paleocene-lower Eocene</td>
<td>38</td>
<td>23</td>
<td>0</td>
<td>1996</td>
</tr>
<tr>
<td>R</td>
<td>G. Umm Mafrud</td>
<td>Paleocene-lower Eocene</td>
<td>27</td>
<td>20</td>
<td>0</td>
<td>1996</td>
</tr>
<tr>
<td>T1</td>
<td>btw. Nuweiba + Taba</td>
<td>Paleocene-lower Eocene</td>
<td>42</td>
<td>21</td>
<td>0</td>
<td>1996</td>
</tr>
<tr>
<td>T2</td>
<td>btw. Nuweiba + Taba</td>
<td>Campanian</td>
<td>88</td>
<td>34</td>
<td>5</td>
<td>1996</td>
</tr>
<tr>
<td>Total</td>
<td>27 sections</td>
<td>?up. Cenom.-mid. Eocene</td>
<td>2395</td>
<td>612</td>
<td>80</td>
<td>1995+96</td>
</tr>
</tbody>
</table>

Tab. 1-1. Sections measured in eastern Sinai during the field seasons in 1995 and 1996. For location map see Fig. 1-1.
2. Geodynamic evolution of Sinai and neighbouring regions in the Mesozoic and Cenozoic

2.1. Summary

During the long geologic history of the Eastern Mediterranean several tectonic processes were active so that the different structures were superimposed on each other (Mart, 1987) (Fig. 2-1, Tab. 2-1). A major element represents a trans-continental strike-slip system which dissects Africa from the northeast to central west and which has been active in phases from the early Paleozoic until today. In the late Triassic / early Jurassic, this zone of weakness as well as other Precambrian lineaments were reactivated when a Turkish-Apulian terrane separated from the Northeast African plate margin and drifted northwards the north. This led to the formation of halfgrabens on the Northeast African passive continental margin which were subsequently transpressively inverted mainly in the late Cretaceous and early Tertiary and resulted in the formation of the Syrian Arc, an intra-continental foldbelt stretching from Northern Egypt to Syria. The compressive stress originated from the collision of the Arabian and Turkish Plates in the area of southern Turkey and neighbouring regions and has been transferred over several hundreds of kilometers to the south within the former halfgrabens systems. Another important geodynamic process is the counter-clockwise rotation of the Arabian Plate relative to Africa which led to the late Oligocene / early Miocene opening of the northern Red Sea, including its northern extension, the Gulf of Suez. The Gulf of Suez soon became a failed rift, probably related to processes in the collision zone of the Arabian and Turkish Plates, which also initiated the opening of the Gulf of Aqaba. The faster northward movement of the Arabian Plate in relation to the Sinai-subplate resulted in strike slip movements at the N-S striking Dead Sea Fault, which leads from Turkey to Sinai and represents a transform-type plate margin. Left-lateral offset is estimated to be 100-110 km.

2.2. Transcontinental Shear System

Since at least the mid-Cambrian, Africa is dissected by a deep-seated, trans-continental shear system, running from central west to northeast Africa with a bended course (Fig. 2-2). Its activity has occurred in phases until Recent. Sense of motion as well as transpressive and transtensive characters changed during the history of the shear system (Neev, 1977, Neev et al., 1985; Keeley, 1994). The 'Pelusium System' as postulated by Neev (1977) and Neev et al. (1985) is composed of the Pelusium Line, the Dead Sea Fault and the Qattara-Eratosthenes Shear. Motion along this old zone of weakness results from differential plate movements between Northwest Africa, the rest of the African Plate and Eurasia, which in turn are associated with large-scale plate movements, namely the stepwise opening of the Atlantic or the collision of the African and Eurasian Plates. The mid-Jurassic opening of the Central Atlantic Ocean, for example, caused Africa to move eastward relative to Eurasia. As a consequence, left-lateral strike-slip movements occurred along the north African faults of the Pelusium System. The late Cretaceous (probably Turonian) opening stage of the Atlantic, in turn, pushed Eurasia to the east relative to Africa causing right-lateral movements in the Pelusium System (Kerdany & Cherif, 1990). The strike-slip system influenced the structure of other regional tectonic elements markedly:

- Orientation of the early Mesozoic northeast African rift grabens and thus the direction of the Syrian Arc anticlines are parallel to the Pelusium strike-slip faults (and other Precambrian sutures), which suggests a reactivation of this old zone of weakness (Neev, 1977, Neev et al., 1985; Barazangi et al., 1994) (see also 2.3.2.).
- The origin of a transtensional basin in northwest Egypt during the Albian has been considered by Keeley (1994) as a direct consequence of transtensional movements along the Pelusium System (see also 2.3.1.)
- Neev (1977) and Neev et al. (1985) interpreted the late Tertiary Dead Sea Fault to be a part of the transcontinental strike-slip system. Therefore the opening of the Red Sea may also be considered as being related to the Pelusium System in a more general view.

2.3. Eastern Mediterranean Basin (Levantine Basin)

Today the main stages of the geodynamic evolution of the Eastern Mediterranean region are well known. The most important models (Sengcir et al., 1984; Robertson & Dixon, 1984; Dercourt et al., 1986, 1993) differ only in a few points concerning temporal and spatial questions of the geodynamic development. A comprehensive discussion of the different models can be found in Robertson et al. (1996). Other summaries of the tectono-sedimentary basin development of the Eastern Mediterranean (=Levantine Basin) have been given by Sestini (1984), Abu-Jaber et al. (1989), Cohen et al. (1990).
<table>
<thead>
<tr>
<th>Geodynamic process</th>
<th>Plate tectonic explanation</th>
<th>Timing</th>
<th>Fig.</th>
<th>Chap.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Transcontinental, deep-seated strike-slip-system (=Pelusium Line = Trans-African Lineament), strike approx. NE-SW, some periods characterized by transpression or transtension related processes:</td>
<td>Mesozoic/Cenozoic. Differential movements between NW Africa, the rest of Africa and Eurasia, caused by large-scale plate processes involving the opening of the Atlantic Ocean and collision of Africa and Eurasia.</td>
<td>Activity in phases from mid Cambrian to Recent (Keeley 1994: 741)</td>
<td>2-2</td>
<td>2.2.</td>
</tr>
<tr>
<td>- Rifting and drifting of the Turkish-Apulian Terrane north of the Egyptian continental mass: Formation of halfgrabens in northern Egypt</td>
<td>Terrane separation is related to the break-up of Pangaea; rifting most probably re-activated an older fault system (Pelusium Line, Pre-Cambrian suture). Formation of a halfgraben system (Ne Egypt, Israel) and a failed rift system (Syria), which lead to increased thickness of sediments on the &quot;unstable shelf&quot;. The &quot;stable shelf&quot; which represents the southern, tectonically stable continental margin, lacks grabens and is characterized by &quot;normal&quot; thickness of sediments.</td>
<td>Late Triassic to &quot;Early Cretaceous&quot; (probably in phases)</td>
<td>2-3</td>
<td>2.3.2.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Early Cretaceous (Albian)</td>
<td>2-4</td>
<td>2.3.1.</td>
</tr>
<tr>
<td>- Formation of a transtensional basin in central northern Egypt (Abu Gharadig Basin, 300 km E of Sinai)</td>
<td>Sinistral movements along the Trans African Lineament (transcontinental fault system, see above)</td>
<td></td>
<td>2-5</td>
<td>2.3.3.</td>
</tr>
<tr>
<td>- Intraplate-folding of the southern Syrian Arc (Sinai-Negev Fold Belt)</td>
<td>Transpressive inversion of halfgrabens, which had been formed during preceding terrane separation. Dextral transpression is related to WNW convergence of Africa towards Eurasia. Closing of the Neotethys along the Bitlis Suture (collision of Arabia and Turkey) and the Zagros Suture (collision of Arabia and Iran) during mid-Eocene to mid Miocene. Convergence between AfroArabia and Eurasia persists until Recent.</td>
<td>Fault has been active in some interval between post-middle Eocene and Pre-early Miocene</td>
<td>2-6</td>
<td>2.3.4.</td>
</tr>
<tr>
<td>- Right-lateral strike-slip fault dissecting central Sinai in an E-W direction including the Themed area</td>
<td>Same mechanism as for intraplate folding (see above), however, lacks compressional component. Fault is interpreted to represent the southernmost fault which had been formed during late Triassic-early Jurassic rifting.</td>
<td></td>
<td>2-7</td>
<td>2.3.5.</td>
</tr>
<tr>
<td>Opening of the northern Red Sea related processes:</td>
<td></td>
<td>Early Miocene (approx. 20 Ma) to Recent</td>
<td>2-8</td>
<td>2.4.1.</td>
</tr>
<tr>
<td>- Movements along the Dead Sea Fault, leads locally to the formation of pull-apart basins as well as folding in the southern Palmyride Fold Belt (Syria)</td>
<td>Continental transform fault which accommodates stress associated with spreading in the Red Sea and collisional processes at the Bitlis Suture (Turkey).</td>
<td></td>
<td>2-7</td>
<td>2.4.2.</td>
</tr>
<tr>
<td>- Neogene folding</td>
<td>Formation of compressional structures related to faulting or due to gravitational gliding of blocks at normal faults (or pull-apart basins) which are associated with the Gulf of Suez and Gulf of Aqaba rifting (independent of Syrian Arc movements).</td>
<td>Folding in the Neogene</td>
<td>2-9</td>
<td>2.4.4.</td>
</tr>
<tr>
<td>- Intrusion of dykes</td>
<td>Volcanism related with the initial opening of the northern Red Sea</td>
<td>Early Miocene (approx. 20 Ma)</td>
<td>2-8</td>
<td>2.4.5.</td>
</tr>
</tbody>
</table>

Tab. 2-1. Overview of the geodynamic history affecting Sinai during the Mesozoic and Cenozoic (references included in text)
Fig. 2.1. Duration of major tectonic processes affecting NE-Africa during the Mesozoic and Cenozoic (after literature data, see references in text). Relative intensity of tectonic process marked by thickness of bar (not indicated for Transafrican Shear System).
2.3.1. Transtensional basin in central north Egypt

Early Cretaceous (Albian) left-lateral transtensive movements along the African trans-continental shear system led to the formation of the E-W striking Abu Gharadig Basin which is located in the deep subsurface about 300 km east of Sinai (Keeley, 1994: 737; Awad, 1984). Timing of the formation of this basin suggests that it has to be treated independently from the rifting and folding on the unstable shelf. It therefore seems possible that similar processes leading to the formation of basins near the Pelusium System may potentially have occurred at many times and at many places, probably independently of regional extensional-compressional trends on the unstable shelf. Such processes must not be neglected in geodynamic reconstructions of the region. Under consideration of these complex relationships, previous tectono-sedimentary models must be tested thoroughly for alternative interpretations, including an involvement of Pelusium strike-slip motions.

2.3.2. Rifting of the Turkish-Apulian terrane

2.3.2.1. Structure and plate tectonic mechanism

The Turkish-Apulian terrane is part of the 'Cimmerian Continent' as postulated by Sengör (1985), and represents a continental strip which separated from Gondwana, drifted to the north and disintegrated. Its fragments collided with Laurasia between the early Carboniferous and early Cretaceous (Sengör, 1979, 1985; Sengör & Yilmaz, 1981; Gealey, 1988; Dercourt et al., 1993). The separation of the Turkish-Apulian terrane from Northeast Africa (Fig. 2-3) is part of the global process of the break-up of Pangaea (Keeley, 1994; Garfunkel, 1995). After the terrane separation, the Eastern Mediterranean (Levant Basin) opened, which Sengör (1985) interprets as the southern branch of the Neotethys. The opening of the Neotethys and the northward drift of the terrane caused a stepwise closing of the Paleotethys, which was completed by mid Jurassic times. The rift was reconstructed as having propagated from west to east and stopped at the Levant coast in Israel. After the rifting, a passive continental margin developed in northern Egypt.

Hirsch et al. (1995a) reject the Neotethys concept with a northern and a southern branch. Based on geologic, paleontologic and geophysical data, these authors favour a model involving a thinned continental crust for the Eastern Mediterranean caused by pre-Late Cretaceous extension and subcrustal magmatic erosion. In this model of intra-continental aborted rifts, the Turkish-Apulian terrane would never have been located at the northern margin of Northeast Africa, but separated from northern Africa by continental crust. This would require a continuation of the Syrian Arc relatively far into the offshore Eastern Mediterranean. The reconstruction of the plate boundary between the African and Turkish Plates remains untouched by the model of Hirsch et al. (1995a), because the terrane model also assumes a collisional zone in the northern Eastern Mediterranean in the Cyprus
Fig. 2-3. Simplified Early Cretaceous paleogeographic map for the Mediterranean and surrounding regions. Note the Turkish-Apulian terrane which had separated from NE Africa-Arabia during the late Triassic-early Jurassic (from Gansser, 1986; modified after Dewey et al., 1973; Moore et al., 1984; and Sengör et al., 1984).

area. The continental-crust model of Hirsch et al. (1995a) is regarded as ‘extreme’ and ‘highly improbable’ by Robertson et al. (1996).

At the northeastern Egyptian continental margin, the late Triassic-early Jurassic rifting resulted in a series of ENE-striking halfgrabens (Fig. 2-4) (Moustafa & Khailil, 1990; Keeley, 1994), which form the structural basis of the ‘unstable shelf’ (after Said, 1962; see 2.3.2.3.). Because of the graben formation, thickness of the Mesozoic sediments is increased in this area (Druckman et al., 1995; Ayyad & Darwish, 1996). The ‘stable shelf’ towards the south remained relatively unaffected by the rifting and contains sedimentary successions with ‘normal’ thicknesses. During rifting, compensation movements took place at the trans-continental shear system (Pelusium Line) and at parallel faults which resulted in the formation of a number of smaller basins (Keeley, 1993).

Palmyride Fold Belt

The Palmyride Fold and Thrust Belt started as a Permian-Triassic failed rift (Fig. 2-5), which was connected to the Levantine rift in Israel and northern Egypt (McBride et al., 1990). Based on modelling of the Bouguer gravity field in Syria, the failed rift and thus the Palmyride Fold Belt reactivated a Proterozoic terrane suture (Best et al., 1990, 1993).

2.3.2.2. Timing of terrane separation and extension

Rifting at the northern margin of Egypt and thus the opening of the Eastern Mediterranean most probably started in the late Triassic / early Jurassic, but according to some sources already in the Permian (see below). A rift event of Karnian / Norian age and first evidence of extension in the Anisian / Ladinian has been reconstructed by Sengör & Yilmaz (1981: 208-9) for south and southeast Turkey which at that time still was part of Gondwana. A late early Jurassic continental shelf-slope-rise triplet has been described from the Levant coast of Israel which is interpreted to have already existed in the earliest Jurassic (Friedman et al., 1971; Goldberg & Friedman, 1974; Bein & Gvirtzman, 1977). Friedman (pers. comm., 1978 in Sengör & Yilmaz, 1981) explains the triplet by a late Triassic rifting event. Laws & Wilson (1996) interpret a similar phase of rifting based on a alkali-magmatic event from north and central Israel. Further to the south, in Sinai, Ginburg & Gvirtzman (1979) found a triplet situation similar to that in Israel. For northern Egypt, Keeley (1993: 219) reconstructed an east-west directed rift propagation. While the rift in Palestine and northeast Egypt had been already active in the Pliensbachian, western Egypt was involved in the extensional movements only from the Bathonian onwards. In the paleogeographic Tethys maps of Dercourt et al. (1993) the Eastern Mediterranean is still closed on the map of the late Norian while the reconstruction for the middle Toarcian already shows the terrane separation.

According to Beydoun (1977) and Garfunkel & Derin (1984), the change from uplift to tectonic subsidence in the Levant region took place already in the Permian and formed the basis for the later rifting. Highly unlikely seems a model of Stampfl & Pillevuit (1993) and Stampfl et al. (1995) who claim an opening of the Eastern Mediterranean already in the late Carboniferous / early Permian. This early opening is also postulated by Guiraud (in press) who reconstructed a break-up of Gondwana at the Afro-Arabian margin during the late Carboniferous. Some authors subdivide the extensional process into different rifting phases. Such phases are reconstructed for the Permian, middle to late Triassic, and early Jurassic (Liassic) (Cohen et al., 1990; Garfunkel, 1995) or in another study (Ben-Avraham & Ginburg, 1990) for the Triassic, Jurassic and early Cretaceous.

Extension in the Levantine region most probably stopped only just before the Senonian compression started. The evidence is described below. This detailed compilation seems necessary because many investigators still base their reconstructions on the models of Beydoun (1977) and Garfunkel & Derin (1984) which postulate that extension terminated at the end of the late Jurassic and was replaced by uplift. From a tectonic sketch in Moustafa & Khailil (1990) it can be interpreted that the lower part of the lower Cretaceous succession in Central Sinai was still influenced by a halfgraben relief, whereas in Northern Sinai the former relief had been already filled and smoothed out (Fig. 2-4). According to Keeley (1994: 736), early Cretaceous sedimentation in northern Egypt was still affected by the halfgrabens. Lower Cretaceous graben-fills have also been described by Chaimov et al. (1992) from the Palmyrides. In a summary of Phanerozoic events in the Levantine Basin, Hirsch et al.
Triassic and Jurassic

Fig. 2-4 (above). Mesozoic and early Cenozoic tectonic development of Sinai. See text for further explanations (from Moustafa & Khalil, 1990).

Early Cretaceous

culmination of Neo-Tethys opening

active rifting

Late Cretaceous

collision along northern Arabian plate margin

reversal of normal fault displacement

NW Aleppo Plateau

Miocene to Present

fold-thrust-type deformation

Fig. 2-5 (left). Mesozoic-Cenozoic tectonic development of the southwestern Palmyride Fold Belt. See text for further explanations) (from Chaimov et al., 1992).
The formation of the Abu Gharadig Basin approximately 300 km west of Sinai in the Albian may provide further evidence that extension in northern Egypt lasted until just before the late Cretaceous inversion and folding of the Syrian Arc (see 2.3.1.). On the basis of the late Jurassic / early Cretaceous extensional movements in the Abu Gharadig and Sirte Basin and of contemporaneous volcanism, Guiraud (in press) and Guiraud & Belion (1995) postulate a new rifting phase for the Eastern Mediterranean during this time.

Using the Ar-Ar method Kilinc et al. (1993) dated alkali basalts from the Negev Desert as early Cretaceous (plateau ages of 120±1.1 Ma and 118.4±1.3 Ma). Alkali magmatites are typically associated with intra-continental rift zones so that the basalts from the Negev Desert provide evidence for activity of the grabens in the early Cretaceous. Basaltic volcanic rocks from the early Cretaceous are also described by Eyal et al. (1995) from southern Israel and similar alkali-magmatic rocks from the late Jurassic - early Cretaceous from NE-Egypt, Sinai, Israel, Lebanon, W-Jordan and NW Syria which according to Laws & Wilson (1996) are also associated with a late Jurassic - early Cretaceous rifting phase in the Eastern Mediterranean.

Hashad (1980: 43-44) and Meneisy (1986) dated an alkali-magmatic event in the southern part of the Eastern Egyptian Desert as late Jurassic / early Cretaceous (150 to 130 Ma or 140 ± 15 Ma). It is unclear whether this volcanism is associated with the rifting in northern Egypt. The question whether a late Jurassic / early Cretaceous hotspot below central north Libya influenced the extensional process in northern Egypt (Morgan, 1982, 1983; Duncan, 1981; Van Houten, 1983) also remains unsolved. A late Jurassic - early Cretaceous heating event is described by Kohn et al. (1995) for southern Israel evidenced by apatite fission track and 40Ar/39Ar K-feldspar data. They explain this heating with the migration of Northeast Africa over the Darfur Hotspot.

In the Palmyrides, first evidence of slightly increased subsidence is reported from the latest Carboniferous or earliest Permian and is interpreted to be associated with rifting (Lovelock, 1984). Chaimov et al. (1992: 707) documented Jurassic and lower Cretaceous rift fills in reflection seismic profiles of the southwestern Palmyrides and thus showed that sedimentation of the early Cretaceous has still been strongly influenced by the extensional structures. It remains unclear, of course, whether extensional movements were still active during the early Cretaceous or whether the structures that had been formed in the Jurassic were just filled. The inverted wedge-shaped halfgrabens described by Chaimov et al. (1992) are 10 km wide, a few 10s of km long and possess graben fills with thicknesses of up to 500 m. Isopach maps indicate elevated thicknesses of the Jurassic and lower Cretaceous deposits within the area of the Palmyrides (200 x 600 km). Taking the data of Chaimov et al. (1992) into account, the model of Lovelock (1984) in which rifting in central Syria was replaced by uplift as early as late Jurassic, seems unlikely.

2.3.2.3. Unstable and Stable Shelves

The tectonic subdivision of Sinai by Said (1962) into a northern ‘unstable shelf’ and a southern ‘stable shelf’ (Fig. 2-6) has been based on apparent differences in structures and sedimentary thickness distribution. Only later, with the establishment of plate tectonics, a reasonable genetic explanation for this tectonic differentiation was found: The unstable shelf represents an area dominated by inverted halfgrabens that were formed when the Turkish-Apulian terrane separated from Northeast Africa in the late Triassic / early Jurassic and inverted from the Senonian onwards (see above). On the stable shelf, no halfgrabens and thus no inversion structures developed.

The boundary between the unstable and stable shelf is formed by the Themed Fault. Moustafa & Khalil (1994) interpret this fault as the southernmost normal fault associated with the Mesozoic rift system in northern Sinai. The strata of the stable shelf in general are flat-lying. At least on the northeastern part of the stable shelf the sedimentation was still influenced by the Syrian Arc deformation. In the late Tertiary, the openings of the Gulf of Suez and the Gulf of Aqaba produced a pronounced new fault pattern on Sinai. Locally, grabens and strike-slip faults of late Tertiary age were formed which complicated the rather simple structural geometries within the Upper Cretaceous-Lower Tertiary units (see 2.4.).

The unstable shelf is characterized by an increased thickness of Phanerozoic sediments and a complex structural setting with anticlines formed by inverted grabens as well as strike-slip faults. The relatively great thickness in this area may be explained by increased subsidence during the Mesozoic rifting phase. The unstable shelf can be subdivided into a basin north of the Sinai Hinge Belt (Fig. 2-6) and an area south of this fault zone. The depocentres of the Triassic, Jurassic, and pre-late Turonian Cretaceous are situated in the northern basin. The basin had an elongated geometry with an ENE-WSW strike and a northeastward dip. Since at least the Turonian, the area was under compressive influence which is documented in the non-homogenous isopach and facies distribution of the Turonian and younger sediments.

The area south of the Sinai Hinge Belt was hardly affected by the Syrian Arc folding, but represented an important depocentre during the late Turonian as documented in isopach and facies maps of the Wata Formation. The different sedimentary developments of the areas north and south of the Sinai Hinge Belt demonstrate that the fault zone must already have existed during the late Cretaceous. Because the tectonic movements in the Sinai...
Hinge belt are of late Tertiary age (see 2.4.3.) it is assumed that this strike-slip fault represents a reactivated normal fault dating from the Mesozoic rift phase. The unstable shelf in Egypt has been tilted westward since at least the late Jurassic, because the early Cretaceous marine ingressions reached the eastern part later than the western part (Kerdany & Cherif, 1990).

[compiled mainly after Kerdany & Cherif (1990) and Cohen et al. (1990)]

### 2.3.3. Syrian-Arc intraplate folding

#### 2.3.3.1. Location and structure

The Syrian Arc (=Levantides Folding Belt [Hirsch et al., 1995b]) consists of the Sinai-Negev Fold Belt (Sinai and Negev-Desert) and the Palmyride Fold Belt (Syria) (Fig. 2-7). Both foldbelts are separated by the NNE-SSW
Fig. 2-7. Generalized tectonic map of the northwestern Arabian Plate and Sinai Subplate. The Syrian Arc consists of the Negev Fold Belt and the Palmyride Fold Belt which are offset by the Dead Sea Transform Fault (from Chaimov et al., 1992).

striking early Tertiary Dead Sea Fault which sinistrally offsets the two units by 100-110 km (Searle, 1994). The name ‘Syrian Arc’ is misleading in the respect that the orogen has nothing to do with an ancient volcanic island arc as the term ‘arc’ would suggest. For historic reasons (Krenkel, 1924, 1925) the term ‘Syrian Arc’ is kept and used in this chapter. Krenkel recognized similarities in strike and structural style of the two foldbelts and called the orogenic system ‘Syrischer Bogen’ (Syrian Arc).

The Sinai-Negev Fold Belt (Fig. 2-7) consists of several belts of domal anticlines which form the highest elevations of the region. Towards the north, the folding structures are covered by Quaternary coastal and shallow marine sediments (Jenkins, 1990: 361). The ENE to NE striking anticlines have doubly plunging fold axes and are highly asymmetric with gentle northwestern (5-20°) and steeper southeastern flanks, which may locally even be overturned. The anticlines form the inventory of a tectonic province in northern Sinai and the Negev Desert which differs markedly from southern Sinai with its flat-lying sediments and its crystalline basement (unstable shelf/stable shelf after Said, 1962). The anticlines have various dimensions. Large anticlines include, for example, Gebel El Maghara (735 m above sea level), Gebel Yelleq (1090 m) and Gebel Halaf (890 m) (Fig. 2-6). Intermediate heights are found in Gebel El Minsherah, Gebel Kherim and Gebel Arif El Naga (Fig. 2-6). In addition, there are many smaller anticlines with diameters up to 2 km. The westernmost structures exposed at the surface include Abu Roash (west of Cairo) and Shabrawet (northern Eastern Desert of Egypt) (Jenkins, 1990; Moustafa & Khalil, 1990, 1995). The westward continuation of this foldbelt may be found in the subsurface of the northern Western Desert (Ayyad & Darwish, 1996).

The Palmyride Fold Belt (Fig. 2-7) has a length of approximately 400 km and, according to reflection seismic data, consists of a 100 km wide zone of intensely deformed Mesozoic and upper Tertiary sediments sandwiched between relatively undeformed subprovinces of the Arabian Platform. Mesozoic and Paleogene sediments in the area of the Palmyrides attain a strongly elevated thickness of at least 5 km (Best et al., 1993). The overall structure of the Palmyrides is characterized by thin-skinned thrust tectonics (Salesl & Séguret, 1994). Folds mapped at the surface can be traced on seismic profiles to a depth of at least 5 km. Similar to the Sinai-Negev Fold Belt the anticlines have an asymmetric structure with a steeper and locally overturned southeastern flank (Chaimov et al., 1992; McBride et al., 1990).

Eyal & Reches (1983) point out that the directions of the anticlinal fold axis (=tectonic macro structures) of the Syrian Arc largely depended on the orientation of pre-existing faults and grabens and thus provide no direct evidence on the paleo-stress field. On the contrary, smaller-scale structures (=tectonic meso structures) such as folds, slickensides, stylolites, dykes etc. are better indicators of the paleo-stress field.
2.3.3.2. Plate tectonic mechanism

The development of the Syrian Arc may be summarized as follows (Fig. 2-4, 2-5):

**Late Triassic / Liassic rifting at the northern margin of the Egyptian continental mass led to the formation of ENE striking halfgrabens which have been dextrally transpressively reactivated since the late Cretaceous because of the starting collision of the African and Eurasian Plates**

The Syrian Arc does not belong to the group of orogens formed by subduction and collision at active plate margins; it rather represents a system of folded (inverted) grabens associated with compressive intraplate deformation. Similar intraplate foldbelts are the Tien Shan in China and the Australian 'Amadeus basin fold and thrust belt'. In northern Europe, the Rotliegend grabens of the Sorgenfrei-Tornquist-zone were inverted during the late Cretaceous and early Tertiary (Michelsen, 1997), more or less contemporaneously to the inversion in northern Sinai. The Atlas mountains in northwest Africa also represent an inversion orogen. Here, a deformational development similar to that in the Syrian Arc is found: 1) pre-Mesozoic fault pattern which was reactivated later, 2) early Mesozoic rifting associated with the break-up of Pangea and 3) inversion of the grabens in phases from the second half of the Mesozoic onwards caused by the convergence of the African and Eurasian Plates (Brede et al., 1992).

Based on the rigidity of the lithosphere, which usually leads to a stress concentration at the plate margins, folding in intraplate orogens needs exceptional crustal properties. In the case of the Syrian Arc, late Triassic / early Jurassic rifting led to the formation of a NE-SW striking graben system in the northeast African and west Arabian regions (see 2.3.2.). From the late Cretaceous onwards, stress initiated in the Bitlis collisional zone in the area of southeastern Turkey (Fig. 2-7), associated with the collision of the African and Eurasian Plates. Under conditions with a homogenous crust, deformation would have been restricted to the plate margins near the Bitlis suture zone. Due to the existence of the zone of weakness along the Levant, the transpressive stress was intra-continentially transferred within the graben system over more than 800 km to the south as far as northern Egypt which led to an inversion of the grabens and thus to the formation of the Syrian Arc (Fig. 2-7).

Subduction in the area of Turkey and Cyprus had been established since the late Cretaceous and the continent collisional stage with the involvement of the Arabian and Turkish-Iranian Plates was reached in the Miocene (Searle, 1994: 1347; Kempler & Garfunkel, 1995). The collisional suture between the Arabian and Turkish Plates is formed by the Bitlis Suture, with its eastward extension, the Zagros Suture, separating the Arabian and Iranian Plates.

Convergence of the two plates has lead to a system of right-lateral shears which reactivated the late Triassic / early Jurassic parallel riffs. Because the direction of convergence forms an angle with the long axis of the the grabens, dextral transpressional movements were established which resulted in the formation of several parallel belts of en-echelon anticlines. The different anticlines represent folded blocks situated between the deep-seated graben-faults. The southeast vergence of the northern Sinai anticlines shows that the stress was oriented from northern directions (Moustafa & Khalil, 1990: 386-389, 1995: 236; Chaimov et al., 1993: 2045).

Salel & Séguret (1994) interpret the northern part of the Syrian Arc, the Palmyride Fold Belt, as a foreland orogenic belt associated with this collisional zone. This implies certain interactions between the Syrian Arc and its southern foreland. It therefore seems necessary to evaluate also the possibility of loading effects for the stable shelf in Sinai. Evidence comes from a late Turonian southward shift of the depocentre to the area south of the Sinai Hinge belt on the southern unstable shelf (Fig. 2-6) (see also 2.3.2.3.).

The convergence between Africa and Eurasia, namely the Arabian and Turkish-Iranian Plates, has lasted until today which is documented by high precision measurements using the Global Positioning System (GPS) (Oral et al., 1995). The Eratosthenes- and Anaximander Seamounts, for example, currently collide with the Cyprus Arc which in contrast to subduction of oceanic crust leads to stress accumulation within the crust. Ben-Avraham & Nur (1986) assumed that this stress, analogous to the situation in the Late Cretaceous, when the Syrian Arc was formed, is released in the early Mesozoic rift grabens of the northeast African continental margin. This also explains the high seismicity which is in excess for a passive continental margin. Since the early Miocene, the compressive stress has also been released and transferred along the NNE-SSW striking Dead Sea Fault and reaches the Syrian Arc system only in part (Lewy et al., 1995: 26); near the intersection of both structural systems, the Syrian Arc structures are transpressively overprinted, caused by the late Tertiary strike-slip movements along the Dead Sea Fault which locally also have a transpressive character. Investigations of the fault pattern in the southwestern Palmyride Fold Belt by McBride et al. (1990: 238) showed that this area is characterized by completely different structures than the rest of the Syrian Arc. Tectonic reconstructions in the southwestern Palmyrides, therefore, must take into account the late Tertiary geodynamic processes and cannot explain the present structures with the Syrian Arc development alone.

The offset of the Sinai-Negev Fold Belt relative to the Palmyride Fold Belt (Fig. 2-7) is predominantly a result of the late Tertiary left-lateral strike-slip movement along the Dead Sea Fault (see 2.4.2.). However, to restore a single straight, linear Syrian Arc Fold Belt, the offset should have been at least 200 km which is not the case. Therefore, it seems likely that both the Sinai-Negev- and the Palmyride Fold Belts were separated right from the
beginning of the folding in the late Cretaceous. The graben system which forms the basis for the direction of the
Syrian Arc might have been offset already during that time by a transform fault.

Seen in a global tectonic context, the origin of the alpidic orogenic system which leads from the west in Gibraltar
to Southeast Asia in the east and which also includes the processes in the Afro-Arabian-Eurasian collisional zone,
is associated with the closure of the Tethys which in turn is a direct consequence of the break-up of Pangaea

2.3.3.3. Timing of deformation

Because of structural and lithological similarities and the more or less corresponding models for the local
deforational development of both parts of the Syrian Arc, it is assumed that folding of the Sinai-Negev Fold
Belt and the Palmyride Fold Belt in general took place contemporaneously and in a similar style (Chaimov et al.,
1992; Shahar, 1994) (Figs. 2-4, 2-5). This is supported by the fact that the compressive stress that is initiated in
the collisional zone in the Turkey area, first has to pass the Palmyride Fold Belt before reaching the Sinai-Negev
Fold Belt. According to this plate tectonic model, the Sinai-Negev- and Palmyride Fold Belt form a single
tectonic system. Discrepancies in the deoromational history of both foldbelts and of different anticlines may be
explained at least in part with different distance to the Afro-Arabian-Eurasian collisional zone. Furthermore,
the complex geometries of the former graben system may have led to different styles of deformation, including
transpression, transtension, pull-apart-structures and pure strike-slip zones. In addition, the late Tertiary
activation of the Dead Sea Fault, which separates the two foldbelts, may have influenced the N-S stress
propagation because it may have served as a stress barrier. Hirsch et al. (1995b) assume that the zone of
maximum deformational intensity during the deoromational history has steadily migrated northward, from the
northern Egyptian craton towards the active collisional zone.

While the detailed deoromational history of the Syrian Arc is still under controversy discussion, a general picture
for the folding processes can be given. First evidence of compression comes from the late Cretaceous. This
datum is in good accordance with the Late Cretaceous development of an ophiolite belt which extends along the
northern margin of the Arabian and northeastern African Plates from Cyprus and southern Turkey to Oman and
indicates an active plate margin for this time period. A compressional regime after the Cenomanian or Turonian
has been reconstructed for Turkey (Sengör & Yilmaz, 1981: 217; Collins & Robertson, 1997) which supports the
collisional model. From the Oman mountains, which are part of the Arabian-Iranian collisional zone, Patton &
O'Conner (1988) describe the uplift of a swell during the Cenomanian and Turonian. Collision of the continental
blocks and thus the closing of the suture along the Bitlis and Zagros suture lines took place only during the
middle Eocene to late Miocene (Yilmaz, 1993; Searle, 1994: 1332; Yigitbas & Yilmaz, 1996). In contrast, Karig
& Kozlu (1990) assume the closing of the Bitlis suture, and hence of the Neo-Tethys, already for the late
Cretaceous. A Quaternary volcanic arc, a high seismicity in the area of the Zagros suture and GPS satellite
measurements indicate that the collision of the Arabian with the Turkish and Iranian Plates is not terminated, yet
(Chaimov et al., 1992: 711-2; Seber et al., 1997). This shows that the collisional processes along the northern
margin of the Arabian Plate which provide the source for the stress in the Syrian Arc have been active from the
late Cretaceous until Recent. It is assumed that the Syrian Arc has been active during the same interval. However,
a large part of the collisional stress has been taken up by the Dead Sea Fault so that tectonic activity in at least
some segments of the Syrian Arc has weakened or even ceased during the late Tertiary. Such a scenario may
especially apply to the southern part of the Syrian Arc, the Sinai-Negev Fold Belt.

Guiraud & Bellion (1995), Guiraud & Bosworth (1996, and in press), and Guiraud (in press) assume that several
short compressional phases were active in Northern Africa and Arabia and they reject the model of a long-lasting
compressional regime. According to these authors, one of the best documented compressional events occurred in
the late Santonian for which evidence was found in Morocco, the Syrian Arc, Oman, the Benue-Tchad-intraplate-
basin and other areas. This event was followed by a Tethyan-wide rifting phase from the Campanian to the
Maastrichtian or Paleocene (e.g. in northern Libya and in the southern Palmyrides), which however on Sinai has
only lead to renewed subsidence after the late Santonian uplift (Guiraud & Bosworth, 1997, in press and pers.
region in Israel and reconstructed a rifting phase for the late Cretaceous which may be equivalent to the
extensional phase postulated by Guiraud & Bosworth (in press). A second compressional phase according to
Guiraud (in press) occurred during the late Maastrichtian. Other short compressional events in Northern Africa
and Arabia were reconstructed for the Aquitain-Burdigal boundary (evidence in the Gulf of Suez and in Oman),
in the Toron (southern Egypt and Israel) and in the early Pleistocene. A similar compressional phase from the
early Pleistocene has also been reported by Sneh (1996) from the Dead Sea Fault.

Sinai-Negev Fold Belt

ANTICLINAL STRUCTURES IN SINAI AND THE NORTHERN EASTERN DESERT OF EGYPT

Until the end of the Cenomanian, no major compressional movements had occurred in northern Egypt as
documented by a more or less uniform thickness of the sediments of this period. A more complex isopach
distribution and pronounced differentiation of the shelf facies in the Turonian of Sinai points to a beginning of
the tectonic activity in the Turonian (Kerdany & Cherif, 1990: 425; Said, 1990b: 445). On seismic profiles, the upper Cretaceous to Oligo-Miocene sediments show clear onlap against the synsedimentary rising anticlinal structures (Ayyad & Darwish, 1996). The most distinct tectonic phase in the late Cretaceous seems to have occurred during the late Coniacian because sediments of this period are reported to be missing in northern Sinai (Lewy, 1975). This may point to a phase of uplift of northern Sinai in the late Coniacian while the central part of Sinai was still covered by the sea. At some locations in northern Sinai, erosion reached as deep as into Turonian horizons. However, it has to be pointed out that the Coniacian erosion may also be at least in part associated with an inter-regional and probably eustatic sea-level lowstand (see chapter 5) so that the tectonic model needs some closer evaluation. Isopach maps of the Campanian-Maastrichtian Sudr Formation indicate a depocentre in central Sinai (Kerdany & Cherif, 1990). The northern margin of the shelf of northern Sinai at that period represented an uplifted area with deposition only in several low-lying areas. Some of the uplifted areas may never have been reached by the sea during that time. In offshore drill holes north of Sinai, for example, Oligocene has been found directly above Turonian (Boughaz 1; Hewaidy, 1993) and Tertiary directly above Jurassic (Slav I) (Kerdany & Cherif, 1990: 425). According to a geological-archaeological study from NW Sinai (Neev & Friedman, 1978), a two-phase tectonic reactivation of the Trans-African Shear (Pelasium Line) took place during the Holocene (2700 b.c. and 15th Century a.c.) which is interpreted to have resulted in further compression within the Syrian Arc structures.

**GEBEL AREF EL NAQA (SINAI)** (Bartov et al., 1980)

First tectonic movements in the area of the Aref El Naqa are dated from the late Triassic / early Jurassic and seem to be a consequence of regional graben formation. Laterally differential subsidence, bounded by faults and flexures, as well as local folding led to partial exposure, non-deposition and minor erosion. During the late Jurassic / early Cretaceous, the area was tilted slightly to the NW, which locally resulted in erosion. During the late Cenomanian and early Turonian, vertical movements occurred that resulted in small lateral fluctuations in facies and thickness. Folding of the anticline took place predominantly between the Coniacian and Maastrichtian with a terminal folding phase after the late Eocene (similar to most of the anticlines in the Negev and in northern Sinai). No evidence of a pre-Senonian elevated structure was found.

**GEBEL YELLEQ (SINAI)** (Moustafa et al., 1991; Moustafa & Khalil, 1995).

First major folding activities started during the late Senonian after deposition of the lower part of the Upper Senonian chalk which resulted in the formation of an intrachalk angular unconformity. Paleocene and lower Eocene rocks are flat-lying or show minor dips in contrast to the folded upper Cretaceous deposits suggesting another angular unconformity between the upper Cretaceous and Paleogene sediments. Similar unconformities are exposed for example at the Mita Pass in NW Sinai (Moustafa & Khalil, 1989) and in the area of Abu Roash, SW of Cairo (Moustafa, 1988; Hamza, 1993). The Gebel Yelleq anticline was probably tectonically overprinted during the early Miocene by oblique-slip faults that are parallel to the Miocene magmatic dykes suggesting a connection with the opening of the Gulf of Suez.

**ANTICLINAL STRUCTURES IN THE NEGEV-DESERT, ISRAEL**

In the Negev Desert, one of the main tectonic phases occurred during the late Turonian to lower Santonian: Short-distance thickness changes in the upper part of the Turonian Nezer Formation and a good correlation between thick intervals and structurally deeper-lying areas indicate an intra-late Turonian initial phase of the Syrian Arc folding. Tectonic activity continued during the Santonian and Campanian and again culminated during the middle Campanian. The Senonian folding phase represents only a part of the evolution of the Syrian Arc, with the deformational history lasting until the Neogene. It is assumed that the Israeli anticlinal structures had reached only half of their modern amplitude until the lower Miocene and that the second half of the compression took place during the Miocene and Pliocene, and probably the Pleistocene (Eyal & Reches, 1983; Braun et al., 1987; Honigstein et al., 1988).

On the basis of subsidence curves derived from drill holes in Israel, Hirsch et al. (1995b) reconstructed several compressional pulses for the Syrian Arc from the Coniacian to lower Miocene with a clear differentiation into initial, main and late phases. From a section in the northern Negev Desert, Zur et al. (1995) described five unconformities for the Turonian to Pleistocene. They correlated these unconformities with the onset of compressional phases.

Shahar (1994: 130-2) subdivided the deformational history of the Sinai-Negev Fold Belt into three phases. Each phase is characterized by a certain tectonic style, and overprinted the structures of the previous phase.

1. **Simple folding, late Turonian to middle Eocene**, 30-50% of the up to 1200 m total amplitude of the different structures; quality and quantity of deposition depending on position within the anticlinal / synclinal systems; basis for estimation formed by isopach and lithological maps; nearly symmetrical folds
2. **Reverse Faulting, late Eocene to mid Miocene**
3. **Differential uplift, late Miocene to Recent**; arching, without steep faulting (see also Shahar, 1995)
17

OFFSHORE, WEST OF ISRAEL

From the offshore area west of Israel, Mart (1994) described early Senonian folding of the Litani anticline. However, shortly after formation of the anticlinal structure, still in the Senonian, extensional processes started and resulted in fast subsidence in the Ptolemaï's-Basin along a transform fault formed during the late Triassic-early Jurassic rifting in the Eastern Mediterranean region. Mart (1994) explained the fast shift from compression to extension by complex collisional and subduction processes in the Afro-Eurasian collisional zone. If the process is dominated by continent-continent or continent-terrane collision, a compressive stress field is postulated for the Eastern Mediterranean. In case of predominant oceanic crust subduction, extension is expected for the northern African Plate. For the late Cretaceous of the Eastern Mediterranean Mart (1994) postulated the presence of short compressive and short extensive deformational phases with relatively long periods of tectonic quiescence in-between. On the basis of analogy with the Palmyride Fold Belt, which reached its deformational climax as late as Oligocene, Hirsch et al. (1995b) expect that the Syrian Arc deformation in the Israeli offshore regions has still been active during the Miocene.

Palmyride Fold Belt (Syria)

Details of the deformational history of the Palmyrides are mostly unknown. On the basis of onlap and downlap structures as well as erosional surfaces in reflection seismic profiles, Chaimov et al. (1992) reconstructed three main deformational phases. The onlap and downlap geometries are interpreted as being related to tectonic events because of several reasons excluding a eustatic origin.

- The earliest uplift took place in the latest Cretaceous (approx. 65 Ma) and resulted in an onlap structure. This phase contributed only a small part to the total deformation.
- The second phase of uplift is dated as middle Eocene and was also manifested by an onlap. Its contribution to the total strain is also interpreted to be limited.
- Uplift activity increased again towards the beginning of the Miocene or already during the middle to late Oligocene. Folding and overthrusting terminated in the Pliocene.

The beginning of this Miocene main folding phase according to Searle (1994) is parallel to
  - Closure of the Bitlis-Zagros Suture
  - Main phase of movements along the Dead Sea Fault and
  - Opening of the northern Red Sea.

On the basis of structural investigations Salel & Seguret (1994) subdivided the deformation of the Palmyrides into two phases. They reconstructed a shortening of 15 km during the late Maastrichtian to late Eocene which lead to the formation of a distinct angular unconformity between the sedimentary complexes of the Cenomanian-Turonian and Maastrichtian-Eocene. For the second compressional phase which has lasted from the late Miocene to Recent, a shortening of 5 km is postulated.

Laramid folding phase

Some authors associate the deformation of the Syrian Arc with the term ‘Laramid folding phase’ which is part of Stille's theory about the global tectonic phases. In this context, the term is used to emphasize the late Cretaceous-early Tertiary deformational period of the Syrian Arc. The term ‘Laramid’ is derived from the orogeny of the eastern Rocky Mountains (including the ‘Laramie Formation’ in Wyoming and Colorado) which was active during the Late Cretaceous to late Paleocene (Bates & Jackson, 1987). According to Stille the Laramid folding phase is part of the Alpidian orogenic phase and includes worldwide all orogenies during the Late Cretaceous-early Tertiary. Attributing an orogeny to a certain folding phase of Stille formerly indicated not only synchronicity but also certain genetic relationships between the orogens of this phase. The replacement of Stille’s orogenic phase theory by the plate tectonic Wilson-cycles requires that the old orogenic phase terms should be avoided and replaced by more neutral age information.

It should be made clear that the observation that certain geodynamic developments occurred synchronously at different places is not questioned by this remark. A late Cretaceous-early Tertiary compressive regime prevailed not only in northeast Africa and northwest Arabia but was also reconstructed for the entire Alpine region (Frisch & Loeschke, 1986) and South and East Asia (Schwan, 1986). Furthermore, also the Jurassic extensional phase of the Alps corresponds well with the development in northeast Africa and northwest Arabia. In northern Europe, lower Permian Rotliegend grabens of the Sorgenfrei-Tornquist-Zone were inverted during the late Cretaceous and early Tertiary, more or less contemporaneously with the folding of the Syrian Arc (Michelsen, 1997; Gabrielsen et al., 1997).

2.3.4. E-W-Strike Slip Fault near Themed

According to Moustafa & Khalil (1994), the late Triassic / early Jurassic rift faults of northern Egypt were reactivated during four tectonic phases (D1 through D4) by right-lateral strike-slip systems. Phase D2 seems to have formed the southernmost Syrian Arc structure - the Themed Fault (Fig. 2-6). Because of the southern,
marginal position of this structure, only strike-slip movements rather than transpressional features have developed at this location.

**Tectonic Phase D1** (compare to detailed description in 2.3.3.)

*Activation of the fault system:* Early late Senonian

*Plate tectonic reason for strike-slip movement:* Collision of the Arabian and Turkish Plates

*Structural Style:* Transpression in Sinai and the Eastern Egyptian Desert, strike-slip faulting in the northern Western Desert of Egypt

**Tectonic Phase D2:** Movements along the Themed Fault (=Ragabet El-Naam Fault)

*Activation of the fault system:* Post-middle Eocene to pre-early Miocene

*Plate tectonic reason for strike-slip movement:* Continuing collision of the Arabian and Turkish Plates

*Structural Style:* Dextral strike-slip faulting, detectable at the Themed Fault and probably at Gebel Areif El Naqa.

The Themed Fault dissects central Sinai in an east-west direction and probably reaches from the eastern margin of the Gulf of Suez to the Dead Sea Fault. Along the fault, the usually flat-lying strata of the Tih Plateau are tightly folded with doubly plunging E-W directed folding axis. It is assumed that the Themed Fault reactivated the southernmost normal fault of the late Triassic / early Jurassic rift system. The fault lies parallel to other strike-slip faults along which the thickness of Jurassic sediments changes drastically. The minimum slip along the Themed Fault is calculated to be 300-750 m. The youngest sediments involved in the structures in the area of the Themed Fault are of middle Eocene age. Movements along the fault must have ended before the early Miocene because an early Miocene dyke intruded the fault which was not offset. This narrows the deformational period of the Themed Fault down to post-middle Eocene to pre-early Miocene.

**Tectonic Phase D3:** only Eastern Desert

*Activation of the fault system:* Late Oligocene to early Miocene

*Plate tectonic reason for strike-slip movement:* Stress compensation for the opening of the Gulf of Suez

*Structural Style:* Transtension; only found in the Eastern Desert of Egypt

**Tectonic Phase D4:**

*Activation of the fault system:* Post-early Miocene to Recent

*Plate tectonic reason for strike-slip movement:* Stress compensation for movements along the Dead Sea Fault

*Structural Style:* Strike-slip faults, movements in the Sinai Hinge Belt.

The Sinai Hinge Belt (Fig. 2-6) marks the boundary between an unfolded zone in the south and a folded unstable zone in the north. Despite structural similarities, the Sinai Hinge Belt and the Themed Fault were active during different periods. The last movements in the Sinai Hinge Belt occurred post-early Miocene because early Miocene dykes are offset by the fault system. Modern seismic activity within the Sinai Hinge Belt indicates that the D4 deformational phase has lasted until today.

(Moustafa & Khalil, 1994)

2.4. Opening of the Red Sea and related processes

The structures associated with the rifting in the Red Sea area dominate the present-day morphology of large parts of Northeast Africa and Arabia. The opening of the Red Sea (2.4.1.) is directly linked to several other tectonic processes, namely to movements along the Dead Sea Fault (2.4.2.), the Sinai Hinge Belt and other Late Tertiary faults (2.4.3.), as well as to folding within the Red Sea fault system (2.4.4.) and intrusion of dykes (2.4.5.).

2.4.1. Opening of the Red Sea, Gulf of Suez and Gulf of Aqaba

The Gulf of Suez has a length of 400 km and a width between 60 and 100 km. The Gulf of Suez represents the northwestern continuation of the Red Sea and separates the African continent from the Sinai Peninsula. The Gulf of Aqaba lies east of Sinai and marks the boundary to the Arabian Plate. A comprehensive summary of the development of the Suez Rift can be found in Patton et al. (1994).

2.4.1.1. Plate tectonic mechanism for the extension

The Red Sea was most probably formed by active rifting (see below) with the initial opening in the southern part. This process led to the counter-clockwise rotation of Arabia relative to Africa and to the separation of both plates creating space for the formation of oceanic crust. The rotational pole is located in the southern central
Mediterranean (Morgan, 1990; Girdler & Underwood, 1985; Girdler, 1991). At the same time, Arabia moved northward relative to Sinai along the left-lateral Dead Sea Fault. As a consequence, the Sinai sub-plate rotated clockwise and was pushed westward which might have partially closed the Gulf of Suez (Jenkins, 1990). Morgan (1990) points out that the Sinai-subplate moved relative to both the Arabian as well as the African Plates which complicates reconstructions of plate movements in the Afro-Arabian region. The amount of extension within the rift system decreases towards the southwest (Gulf of Aden) and towards the north (Gulf of Suez and Gulf of Aqaba), which is associated with the position of the rotational pole (McKenzie et al., 1970; Freund, 1970; Le Pichon & Francheteau, 1978). While the Gulf of Aden, the Red Sea, and the Dead Sea Fault represent the most active parts of the system today, the Gulf of Suez and the East African Rift are considered as more inactive elements. The latter meet the actively opening system at the Sinai and Afar triple junctions. The Gulf of Suez, which is relatively shallow, must probably be interpreted as a "failed rift" (Courtillot et al., 1987; Searle, 1994: 1336).

In general, two end members of rift processes can be defined (Sengör & Burke, 1978; Baker & Morgan, 1981):
1. **active rifting**
   - active asthenospheric upwelling; uplift and volcanism precede the establishment of extensional faults.
2. **passive rifting**
   - results from stress at plate margins; establishment of extensional faults precede the onset of volcanism.

Evidence of active rifting in the Red Sea is provided by heat flux data from the Red Sea indicating higher crustal temperatures than would be expected in a setting with pure passive extension of the lithospheric crust (Morgan et al., 1985; Morgan, 1990). Furthermore, the graben shoulders experienced a strong uplift (Steckler et al., 1994), which under normal conditions does not occur with passive rifting.

Evidence against an active rifting process is provided by the fact that uplift did not precede extension (thermal doming) but occurred almost contemporaneously with the main extensional phase (Steckler, 1985). The absence of Oligocene sediments in the area of the Gulf of Suez which in the literature has often been considered as evidence for doming, is interpreted by Steckler (1985) as non-deposition during a major regression. Steckler (1985) explains the strong uplift of the graben shoulder of the Gulf of Suez, which is too strong for passive rifting, with the activity of smaller secondary convection cells within the rift induced by high thermal gradients during the extensional process.

The formation of oceanic crust at a fully developed mid-oceanic-spreading ridge within the Red Sea basin seems unreasonable because a morphological ridge is not detectable in the modern bathymetry (Morgan, 1990). Exceptions are the southern Red Sea and the Gulf of Aden for which an origin by classical sea floor spreading is assumed (Cochran, 1981, 1982, 1983; Girdler & Styles, 1982). For the main part of the Red Sea system, it is interpreted that the subsided continental blocks within the rift graben as well as the graben margins were intruded by numerous dykes, which led to an oceanisation of the crust in the Red Sea (Fig. 2-8). This process resulted in the formation of a hybrid-continental crust, which today is characterized by physical properties (e.g. seismic velocities etc.) ranging between such of pure continental and pure oceanic crustal types. In addition, the dyke intrusions created a magnetic anomaly pattern in the crust of the northern Red Sea which, with some exceptions, does not have such a classical linear distribution as for example in the Atlantic (Girdler, 1985; Baldridge et al., 1991).

---

**Fig. 2-8.** Map of the Red Sea/Gulf of Suez in the early stage of formation (about 20 Ma). Note that the Gulf of Aqaba is still closed. Dykes trend parallel to the Gulf of Suez rift axis (from Baldridge et al., 1991).
2.4.1.2. Timing of extensional movements

Rifting and the separation of Arabia from Africa (Oligocene) in the southern Red Sea started at about 30 Ma and according to most authors commenced in the northern Red Sea at about 20 Ma (early Miocene), based on radiometric datings of basaltic dykes which intruded during the initial rifting stage (see 2.4.5.) (Baldridge et al., 1991 and references therein).

Rifting in the Gulf of Suez started contemporaneously with that of the northern Red Sea. This assumption is supported by apatite fission-track analyses (Kohn & Eyal, 1981) which allowed the reconstruction of the uplift history of Sinai with the beginning of rifting at around 22 Ma. According to these investigations, uplift of the Sinai-Peninsula started at 26.6 ± 3 Ma (late Oligocene), which seems to exclude an earlier rifting in the Gulf of Suez. In contrast, Harland et al. (1989) postulate first movements associated with rifting already during the latest Eocene (35 Ma).

Apart from intrusion of dykes in the area of the Gulf of Suez, extensional block tectonics also took place. Shortly after these processes had started, tectonic processes within the Arabian-Eurasian collisional zone changed the regional stress field in the region of the northern Red Sea. As a consequence, rifting activity switched from the Gulf of Suez to the Gulf of Aqaba. The Gulf of Suez turned into a failed rift (Lyakhovsky et al., 1994: 31, 41; Searle, 1994) and was even inverted in part (Knott et al., 1995). Girdler (1991) reconstructed two main extensional phases for the northern Red Sea. The first phase is assumed to have taken place between 24-16 Ma, a second phase is postulated to have started at 5 Ma and to be still active today. Differentiation into two extensional phases is explained by Krauss & Baabbad (1987) as a consequence of processes in the collisional zone between the Arabian and Turkish/Iranian Plates along the Bitlis- and Zagros-Sutures, which influenced the plate movements in the area of the Red Sea and thus the opening process of this ocean basin.

The beginning of Red Sea rifting in the Egyptian region during the Eocene is postulated by Morgan (1990: 104) and Bunter (1982: 22). According to Girdler (1991) a first continental rifting phase with a local development of a proto-Red Sea Basin in the Red Sea occurred in the Eocene. Only few authors assume an even earlier onset of rifting in the northern Red Sea during the Paleocene and Cretaceous. Their evidence comes from a late Cretaceous alkali-magmatism interpreted to be associated with doming of the crust in the area of the Red Sea (Ragab & El-Kalilouby, 1992). The presence of a doming stage would indicate 'active rifting' and would have preceded an extensional stage. Reconstruction of sedimentary transport directions from the area of the Red Sea seems to support the doming model (Ward & McDonald, 1979). In contrast, Meneisy (1986) identified a late Eocene / early Oligocene magmatic event (40 +/- 10 Ma), which he interprets as being associated with doming and extension, and a second late Oligocene / early Miocene magmatic event (24 +/- 2 Ma), which he assumes to represent the initial opening of the Red Sea.

Strong variations in thickness within short lateral distances in Upper Paleocene and Eocene sediments made Strougo (1986) postulate a late Paleocene rifting event associated with the formation of the Red Sea (velascoensis Event). Strougo (1986) described abrupt local shallowing of the depositional regime as well as coarse bioclastic mass flow sediments from an area reaching from the Egyptian Western Desert to the Gulf of Suez / SW Sinai, and (mis-) interpreted (see chapter 9) them as a consequence of active block faulting and formation of a strong extensional relief. According to Strougo (1986) the pre-Upper Paleocene sediments indicate rather constant depositional conditions. A similar model was proposed by Jordi (1984), who interpreted the formation of NW trending horst blocks towards the latest Cretaceous and in the Paleocene as a consequence of the Gulf of Suez rifting. According to Keeley (1994) the Gulf of Suez has been a zone of weakness since at least the Carboniferous and was only reactivated later by the rifting of the Red Sea Rift system. However, facies and isopach maps for the Cenomanian and Lower Senonian which also cover the marginal and central parts of the Suez Rift do not provide any evidence of the existence of a rift structure during this time (Garfunkel & Bartov, 1977).

2.4.2. Movements and structures at the Dead Sea Fault

The Dead Sea Fault (~Levant Shear) extends from the Taurus Mountains in Turkey as a strike-slip fault to the Gulf of Aqaba where it continues as a rift further south. The Dead Sea Fault and the Gulf of Aqaba rift together have been termed 'Dead Sea Rift' (Lyakhovsky et al., 1994), which forms a transform-type plate boundary between the Sinia subplate and the Arabian Plate. The horizontal displacements are the result of differential movements in the northward drift of Africa and Arabia. Arabia is moving relatively faster by approximately 15 mm/year, which is accommodated by left-lateral offset along the Dead Sea Fault. Stress in the Dead Sea Rift is derived from two different transform systems, which are active contemporaneously and unify in the Dead Sea Rift. The first transform fault is based in the northern Red Sea and is propagating through the Gulf of Aqaba towards the north. The second transform fault begins in the Bitlis collisional zone and propagates to the south (Lyakhovsky et al., 1994; Dalkal et al., 1990).

The Dead Sea strike-slip fault has been active since the late Oligocene / early Miocene (about 20 Ma) to Recent. The offset which can be proven today, has been reconstructed by most authors as 100-110 km and is interpreted after stratigraphic, lithologic, structural and geophysical data (Quennell, 1958, 1984; Freund et al., 1970;
Garfunkel, 1981; Garfunkel et al., 1981; Hatcher et al., 1981; Girdler, 1985; Sneh, 1996). It is assumed that the horizontal displacements took place during two phases: 60-65 km offset are interpreted as being pre-Miocene, while the remaining 40-45 km are associated with movements since the Miocene (Freund et al., 1970; Quennell, 1984, Girdler, 1985). Darkal et al. (1990) have correlated these two deformational phases with the main spreading phases (24-16 Ma and 5 Ma to Recent) which are reconstructed for the Red Sea. In contrast, Sneh (1996) postulated, on the basis of stratigraphic investigations and facies reconstructions that almost the entire offset of about 100 km was achieved from the late Oligocene to the early Miocene and that younger extensional and compressional deformational phases caused mainly vertical movements. The offset of 100-110 km has been reconstructed only for the southern part of the Dead Sea Fault (up to the Jordan Graben). In the northern part of the fault zone only 20-30 km offset have been proven. The remaining strain may have been accommodated by conjugated shear systems, which are partly located offshore in the Eastern Mediterranean, or within the southern Palmyride Fold Belt.

Shahar (1994) assumes for the entire Dead Sea Fault an offset of 45 km based on well data. However, this model may be questioned because of the strong evidence of the 100-110 km offset as provided by many other authors.

2.4.3. Movements along the Sinai Hinge Belt and late Tertiary faults

The Sinai Hinge Belt consists of two parallel ENE-WSW trending strike-slip faults. It is located in Central Sinai and marks the boundary between (mostly) unfolded crust in the south and a folded area in the north (Moustafa & Khalil, 1994). However, the Sinai Hinge Belt does not represent the boundary between the unstable and stable shelf, which is located further to the south at the Thumed Fault (Kerdany & Cherif, 1990).

The right-lateral strike-slip faults in the Sinai Hinge Belt (Shata, 1959) (=Minsherah-Abu Kandu Shear Zone, =Central Sinai-Negev Shear Zone (Bartov, 1974)] (Fig. 2-6) were activated in post-early Miocene to Recent times (deformational phase D4 in Moustafa & Khalil, 1994, see 2.3.4.). The last movements in the Sinai Hinge Belt occurred post-early Miocene because early Miocene dykes are offset by the fault system. Modern seismic activity within the Sinai Hinge Belt indicates that deformation within this zone has lasted until today. This activity has to be considered as a compensation for movements in the Dead Sea Fault (Moustafa & Khalil, 1994).

On Sinai, numerous faults are developed that accommodate movements from the Dead Sea Fault. For example, in southwest Sinai a belt of mainly subparallel N-S and NE-SW striking faults can be found which have a cumulative offset estimated to be 24 km (Eyal et al., 1981). Interestingly, not all faults are subparallel to the Dead Sea Fault or are of sinistral character as the Dead Sea Fault, because they often reactivated Precambrian zones of weakness within the crystalline basement (Frei & Freund, 1990). Faults parallel to the Gulf of Aqaba strike (NE-SW) or to the Gulf of Suez (Erithrean-) strike (NW-SE) are most probably associated with the rifting (Kerdany & Cherif, 1990).

A fault that has only been found in the Egyptian Eastern Desert was active between the late Oligocene and early Miocene (deformational phase D3 in Moustafa & Khalil, 1994). The transtensive movements may be explained as compensation for the opening of the Gulf of Suez. (see 2.4.1.) (Moustafa & Khalil, 1994).

2.4.4. Neogene folds in the Red Sea fault systems

During the late Tertiary, folding structures developed south of the Syrian Arc deformational front in central and southern Sinai. Their origin differs markedly from the development of the northern Sinai anticlines. These early Tertiary folds were formed in association with strike-slip faulting in the Dead Sea Fault / Red Sea systems and are arranged in en-echelon patterns. Examples of these folds have been described by Abdel-Khalek et al. (1992, 1993) from SE-Sinai and by Moustafa & El-Raey (1993) and Moustafa (1993) from SW-Sinai. Some of these late Tertiary folds may also have been formed by tilted blocks which gravitationally slid into grabens or pull-apart basins leading to compressional structures within the gliding blocks and the substratum (Fig. 2-9) (Abdel-Khalek et al. (1992, 1993: 434).

2.4.5. Intrusion of dykes

During the initial phase of rifting, numerous basaltic dykes intruded the whole area of the Red Sea and the Gulf of Suez with orientations parallel (and subordinately also obliquely) to the long axis of the rifts. The intrusions most probably were implaced during a single event at about 20 Ma (early Miocene) because all dykes - regardless of their distance to the central rift zone - are of almost the same age so that a diffuse intrusion mechanism has to be assumed (see 2.4.1.1.) (Baldridge et al., 1991). On Sinai, the dykes can be found in the Precambrian basement as well as in the Phanerozoic sediments. It is assumed that the basaltic magmas originate from the lithospheric mantle (Baldridge et al., 1991; Eyal et al., 1981).
2.5. Importance of the tectonic processes for Cretaceous-Paleogene basin analysis in Sinai

Sea-level changes reconstructed from sediments in tectonically active regions usually include a eustatic and a tectonic component. A differentiation between these two components for the Areif El Naqa anticline within the Syrian Arc Fold Belt in northern Sinai is attempted in chapter 9. In order to gain a better understanding of the tectonic processes potentially active during this time on Sinai, not only the Syrian Arc compression had to be checked for its regional and temporal activity, but all other tectonic processes from the Mesozoic and Cenozoic [namely the Transafrican Lineament (2.2.), Mesozoic rift grabens (2.3.2.) and late Tertiary Red Sea rift system (2.4.)] as well. Marginal and seemingly less important processes in terms of regional and temporal distribution received special attention in this review chapter because it was necessary to evaluate whether they ‘still’ or ‘already’ affect the sedimentary processes of this time interval.

The literature survey has shown, for example, that deposition of the oldest sediments studied in this project (lower Turonian) was no longer influenced by a half-graben-system relief in Northern Sinai, whereas this cannot be assumed for the underlying Lower Cretaceous in general, for which several authors postulate the presence of such a relief (see 2.3.2.2.). Another geodynamically important item is to evaluate the possibility of an early rifting of the northern Red Sea and its northern extensions. Strougo (1986), for example, postulates synsedimentary block faulting during the latest Paleocene for the region from the Western Desert to the Gulf of Suez and interprets it as being related to an early stage of Red Sea opening (‘velascoensis event’). However, his ‘evidence’ and an alternative view is shortly described in chapter 9.

A look at the larger-scale geodynamic processes shows that the late Cretaceous - early Tertiary folding in the Syrian Arc is a direct consequence of the collision between Eurasia and Africa and that the continent-continent collisional stage was reached only in the early Tertiary, being much later than the deposition of the youngest sediments investigated in this study. The compilation of data from the entire collisional history until Recent aims at a better and more comprehensive understanding of the compressional processes. Chaimov et al. (1992: 712-3) point out that the tectonic development of the Syrian Arc most probably was synchronous to the deformational processes in the Bitlis-Zagros suture zone. A detailed analysis of the geodynamic intraplate development in the less deformed Syrian Arc may provide a better understanding of the intensely deformed, and therefore more complicated to reconstruct, inter-plate collisional zone.
3. Regional sedimentary setting

3.1. Stratigraphic Overview

The Upper Cretaceous to Early Tertiary sedimentary succession in Eastern Sinai is dominated by shallow marine to outer shelf carbonates with rare, but from a genetic point of view important, intercalations of siliciclastics. Detailed descriptions of these successions can be found in chapters 5, 7 and 9. It has to be pointed out that the reconstructions in the present study are predominantly based on a (multi-)biostratigraphic frame and that correlations of the regionally established lithostratigraphic units provided only some additional data. Nevertheless, all measured sections were lithostratigraphically subdivided and formations genetically interpreted where possible, in order to facilitate comparisons with literature data from the region which in many cases is based on lithostratigraphy (e.g. isopach maps; Bartov & Steinitz, 1977).

Owing to the intermediate position between mainland Egypt and Israel, two different lithostratigraphic systems have been used for the Upper Cretaceous-Lower Tertiary in Sinai. While the Egyptian formations are mainly based on Ghorab (1961), the Israeli nomenclature was developed by Flexer (1968) and Bartov et al. (1972). Correlations of the different schemes (Fig. 5-2) are included for example in Kerdany & Cherif (1990), Kuss (1992) and Kora & Genedi (1995). In the following, a short characterization of the different formations in the study is given, while detailed lithological descriptions and remarks on differentiation of the formations in central east Sinai can be found, for example, in Bartov et al. (1972), Ziko et al. (1993) and Kora & Genedi (1995).

The Abu Qada Fm. (=Ora Shales in the Israeli nomenclature) consists of mainly pelitic lithologies and is attributed to the late Cenomanian to early Turonian. These soft lithologies are overlain by hard limestones of the late Turonian Wata Fm. (=Gerofit Fm.). The Coniacian-Santonian succession is included in the Matulla Fm. (=Zihor Fm. and lower part of Sayyarim Fm.) consisting mainly of calcareous siliciclastics, carbonates and inner shelf shales. Towards the north, the siliciclastics are replaced by carbonates which are grouped into the Themed Fm. in this area. These units are overlain by the Campanian-Maastrichtian Sudr Chalk (upper Sayyarim Fm. and Ghareb Fm.) which is composed of chalks, cherts and limestones. The Paleocene to lower Eocene succession consists predominantly of greenish marls which are termed ‘Esna Shales’ in Sinai and Taqiye Fm. in the Negev. This soft unit is overlain by hard dolomitic carbonates of the middle Eocene Thebes Fm. (=Mor Fm. of the Avedat Group in the Negev Desert).

3.2. Hydrocarbon provinces in Sinai

Petroleum was the only mineral raw material produced in Egypt in a considerable amount in 1989 (Dolley, 1989). In 1992, the approximately 100 fields in Egypt produced 45 million tons of petroleum and 9.5 billion m$^3$ of gas. This places Egypt in an intermediate rank among the hydrocarbon producing countries (Sestini, 1995: 57). Sinai shares parts of two hydrocarbon provinces, namely the

- Gulf of Suez rift basin (SW-Sinai) and the

3.2.1. Gulf of Suez rift basin

The Gulf of Suez Basin is the main petroleum province of Egypt (Figs. 3-1, 3-2) and holds a 7th rank in worldwide production from rift basins (Clifford, 1986; Schlumberger, 1995). The NNW-striking rift has a length of 320 km and forms the northwestern continuation of the Red Sea. The width of the rift varies between 50 and 90 km, while only 20-30 km are covered by shallow marine waters with a maximum depth of 40-60 m (Sestini, 1995).

The formation of the hydrocarbon-rich reservoirs in the Gulf of Suez is connected to late Oligocene / early Miocene rifting which led to intense block tectonics within the Paleozoic to Eocene pre-rift succession. During this extensional period, a thick Miocene syn-rift series with excellent source, reservoir and seal qualities was deposited. Similar conditions can be also found for some horizons of the pre-Miocene strata. The development of steep normal faults since the Miocene has produced complex tectonic geometries suitable to serve as structural traps (El Ayouti, 1990; Salah & Alsharhan, 1996). Furthermore, the strong horst-graben relief has led to the formation of stratigraphic traps within the Miocene succession.
Fig. 3-1. The Eastern Mediterranean region showing oil and gas fields. Hachured area indicates Mesozoic Basin (from May, 1991).

Fig. 3-2. Map of petroliferous basins of Egypt showing oil and gas fields and discoveries in the Western Desert, the Nile Delta and Sinai. In the Gulf of Suez oil fields producing >5000 bopd are indicated (from Sestini, 1995).
Main important parameters in the Gulf of Suez hydrocarbon play:

Source rocks
For many years the middle Miocene Globigerina marls and shales of the basinal facies were interpreted to be the only source for the oil in the Gulf of Suez basin. Today, however, it is assumed for several areas that the main source is constituted by pre-Miocene rocks with an only small contribution by the Miocene Globigerina marls. Potential pre-Miocene source rocks include Upper Senonian to Eocene limestones, chalks and marls (Brown Limestone, Duwi Fm., Sudr Fm., Esna Shale, Thebes Limestone). Lower Cretaceous and Palaeozoic shales also possess source qualities, but they are mainly gas-prone (Sestini, 1995; Salah, 1989).

Reservoir rocks
Within the Gulf of Suez, there are several horizons with good reservoir qualities. The main reserves, however, are stored in massive Cambrian to Lower Cretaceous Nubian sandstones. Other reservoirs of local importance are fractured crystalline basement, lense-shaped sandstones from the Cenomanian-Turonian and Coniacian-Santonian as well as fractured, cavernous limestones from the Upper Cretaceous to Eocene. In some fields Miocene syn-rift reservoirs are developed (Sestini, 1995; Enani et al., 1994).

Seals
Horizons with seal potential can be found in the whole lithological succession in the Gulf of Suez region. Especially well developed seals, however, are developed within the Cretaceous and the Miocene intervals containing shales, marls, dense limestones and evaporites (El Ayouty, 1990; Salah & Alsharhan, 1996). The Miocene evaporites form the most effective seals (Sestini, 1995).

Traps
- Pure structural traps are developed in fields of pre-Miocene rocks. Important structural elements are block faulting, faults and fault-related flexures.
- Fields located within Miocene rock units are characterized by a combination of structural and stratigraphic trap. The well-developed horst-graben relief led to strong lateral variations in facies and thickness of the Miocene syn-rift sediments (e.g. wedging out of strata, local reefs etc.) (El Ayouty, 1990; Sestini, 1995). In combination with the structural elements of the rift, numerous, in part complex traps were formed.
- Until now, no oil has been found in the anticlinal structures of the Syrian Arc in the northern Gulf of Suez (compare to section below about the Mesozoic Basin in the Eastern Mediterranean) (El Ayouty, 1990: 569).

Maturation
Rapid tectonic subsidence represents the driving force for maturation of hydrocarbons in the Gulf of Suez (Sestini, 1995). Most investigators postulate that the greatest part of the oil was formed during the period of 8 to 4 million years ago. In contrast, Feinstein et al. (1995) found that maximum heating of the rock units has only been reached in the Holocene, at least at the eastern margin of the Gulf of Suez Rift, resulting in a late to sub-Recent formation of the Gulf-of-Suez oil. There interpretation is based on studies of the maturity of organic matter, apatite-fission track investigations and temperatures in drill holes. According to Salah & Alsharhan (1996: 115), some of the rift blocks have subsided to a depth at which the pre-Miocene pre-rift sediments and the syn-rift sediments have reached the gas-window.

3.2.2. Mesozoic Basin in the Eastern Mediterranean

The Mesozoic basin in the Eastern Mediterranean has an extension of more than 2000 km and reaches from the Western Egyptian Desert, Northern Egypt, Israel, the Lebanon, Jordan, Syria and southeast Turkey to northern Iraq (Fig. 3-1). Several subbasins are developed (Keeley, 1994). The following description concentrates on the southern part of the basin, although many statements are valid for the entire basin.

During repeated rifting phases from the Permian (Triassic) to the Lower Cretaceous, a mosaic of horst-graben systems was formed in the southern Eastern Mediterranean region. After the main subsidence peaks in the grabens had terminated, depositional sag basins developed. The direction of tilt of the major regional blocks changed several times during their evolution which may be associated with changes in the inter-regional stress field in the collisional zone of the Afroarabian and Eurasian Plates. From the Late Cretaceous onwards Syrian-Arc compression led to an inversion of the graben systems. Asymmetric fault-fold sag basins formed at the flanks of the anticlinal inversion structures (Cohen et al., 1990; see also chap. 2).

The intense rift and inversion tectonics in combination with strong eustatic sea level changes led to the development of strong lateral and vertical facies variations with differentiation into several subbasins. This resulted in a large variety of source, reservoir, and sealing lithologies, which in some places were formed in close proximity so that stratigraphic traps evolved. Furthermore, the extensional and compressional movements are responsible for the formation of a multitude of potential tectonic traps.
Good summaries of the basin development in the southern Eastern Mediterranean region with an emphasis on hydrocarbon genesis and distribution can be found in Cohen et al. (1990), May (1991), Zaghoul & Khidr (1992), Ayyad & Darwish (1996) and Alsharhan & Salah (1996). Overviews of the petroleum geology in Egypt with chapters on the Gulf of Suez and Northern Sinai were published by El Ayouty (1990) and Sestini (1995).

Until 1995, 43 petroleum exploration wells have been drilled in Northern Sinai, most of them 'dry holes'. The area is classified as relatively little explored (Sestini, 1995: 83). The high percentage of 'dry holes' stopped all major exploration campaigns in the formerly highly interesting Northern Sinai basin. In a few cases it was found that hydrocarbons which formerly had been accumulated, have been degraded by bacteria from meteoric waters (oral communication, AGIP, 1996). Currently, Triassic plays are interpreted to be the economically most interesting plays because oil and gas has been found in similar geologic settings in Israel and Syria. It seems that the most-promising trap-seal combinations are developed in the coastal or near-offshore zone of Northern Sinai (Sestini, 1995: 83).

The major play parameters of the Mesozoic Basin in the Eastern Mediterranean [compiled predominantly after Cohen et al. (1990), May (1991) and Alsharhan & Salah (1996)]:

Source rocks
During most of the Upper Cretaceous and Lower Tertiary chalks, marls and shales with high degrees of organic carbon were deposited in the entire basin. They form the basis of the largest part of the hydrocarbon reserves in the Eastern Mediterranean basin. Additional bitumen-rich basinal sediments can be found in some of the older stratigraphic units, as for example the fine-grained facies of the Mesozoic grabens and sag basins. In contrast, Alsharhan & Salah (1996) [based on Tammam (1994)] consider lower Cretaceous shales as the most prolific source rocks in the offshore part of northern Sinai while Jurassic, Oligocene and Miocene shales may gain local importance. It is interesting to note that Alsharhan & Salah (1996) do not describe a source rock potential for the Upper Cretaceous-Lower Tertiary succession.

Reservoir rocks
The clastic facies at the margins of the Mesozoic grabens and sag basins represent excellent reservoirs. Mesozoic siliciclastic fans from the Arabian Shield and porous and fractured carbonates also possess a high reservoir potential (e.g. the fractured Thebes Limestone, El Ayouty, 1990:597). Carbonates of the Upper Turonian Wata Fm. and the Coniacian-Santonian Matulla (Themed) Fm., which are also studied in this thesis, provide the main reservoir in the Sadot gas field (NE Sinai) and the Raad discovery (Alsharhan & Salah, 1996). Of minor importance are Paleozoic sandstones and Triassic carbonates which form small reservoirs in the offshore of northern Sinai (Alsharhan & Salah, 1996).

Seals
The fine-grained deposits of the pelagic graben- and sag-basin facies provide good seals. In some of the sag basins, evaporitic horizons with excellent sealing qualities are developed. Carnian anhydrite, for example, is found at the Gebel Areif El Naqa anticline in NE-Sinai. Other seals are the fine-grained sediments of the outer shelf and non-fractured carbonates, for example the Paleocene Esna Shale and the Eocene carbonates of the Thebes Fm. which seal Upper Cretaceous reservoirs.

Traps
Because of the intense rift- and inversion tectonics, several different tectonic trap structures are developed (e.g. fault blocks, asymmetric anticlinal structures, traps at faults). In addition, several stratigraphic traps are developed, for example lens-like sand bodies in shales, reefs and unconformities.

Repeated tilting of regional blocks in the Eastern Mediterranean led to the formation of large sedimentary fans. During the Early Cretaceous, for example, the Levant block was tilted basinwards towards the north which resulted in a favourable seal-reservoir combination. In this case, siliciclastics with a high reservoir potential were shed from the Arabian Shield and interfingered with shaley, bitumen-rich sediments of the basinal facies. A similar favourable situation is interpreted for the mid Triassic in the southeast of the Negev plateau.

Maturation
Sufficient maturities are only developed at places where the source rocks reached relatively great depths. This for example can be found in the Mediterranean offshore-area where the Mesozoic and Early Tertiary rock units are overlain by several thousand meters of Neogene sediments. According to Alsharhan & Salah (1996) the main source kitchens for the Jurassic and Lower Cretaceous in Northern Sinai are the East Maghara, North Maghara, Northeast Sinai and the Tineh-Mango troughs. The Oligocene and Miocene source kitchens in north Sinai are the Wakar-Port Fouad and Tineh-Mango troughs. The Oligocene source rocks are mature in the northern part of the North Maghara trough (Alsharhan & Salah, 1996).
4. General concepts

4.1. Sequence stratigraphic concept

The sequence stratigraphic concept was originally developed for seismic profiles. Sequences and systems tracts were defined by typical stratal geometries such as onlap, offlap and downlap. These large-scale structures, however, are often not resolvable in outcrops or cores so that in such cases other criteria for the recognition of the sequence stratigraphic units have to be used. The wide variety of methodological approaches and directions of emphasis led an extensive sequence stratigraphic nomenclature. The non-standardized usage of the often redundant or ambiguous terms caused misunderstandings in many cases (Posamentier & James, 1993; Van Wagoner, 1995: xii). In addition, many workers seem to use the sequence stratigraphic concepts without having fully understood them (Posamentier & James, 1993).

Sequence stratigraphy represents a tool for the processing and interpretation of geological data rather than a rigid, unmodifiable law. It is highly unlikely that the geologic situations found in nature, exactly correspond to the theoretical, simplified sequence stratigraphic models. Therefore, every sequence stratigraphic study has to consider the different regional geological situations and their effects on the tectono-sedimentary system before applying the sequence stratigraphic concepts.

Sequence stratigraphy can be used in two ways: Age models are constructed by using sequence stratigraphic surfaces as datum lines for the lateral age correlation. In a second application, depositional cyclicity is used for the prediction of the lateral lithological development (Posamentier & James, 1993).

General definitions


sequence stratigraphy: Sequence stratigraphy subdivides the deposits of a sedimentary basin into genetically linked, repetitive sedimentary packages (depositional sequences) within a chronostratigraphic frame (Van Wagoner et al., 1988; Posamentier et al., 1988).

sequence (depositional sequence): Relatively conformable succession of genetically linked strata which is bounded by an unconformity and its correlative conformity (Mitchum, 1977). Parasequences and parasequence sets form the building blocks of a sequence (Van Wagoner et al., 1990). Sequences are subdivided into systems tracts, namely (from base to top) the lowstand-, transgressive- and highstand- / shelf margin systems tract (Vail et al., 1991). Sequences are by definition independent of time and space (Posamentier & James, 1993: 9).

parasequence (ps): Relatively conformable shallowing-upward succession of genetically linked strata which is bounded by a flooding surface or its correlative horizon (Van Wagoner, 1985; Van Wagoner et al., 1987, 1988, 1990). The ps-definition is largely independent of the sea-level development (with the exception of the required flooding surface), stratal geometries and other interpretative sequence stratigraphic aspects (Posamentier & James, 1993).

parasequence-set: Succession of genetically linked parasequences with a typical stacking pattern (e.g. prograding, retrograding / backstepping) which is bounded by a flooding surface or its correlative horizon (Van Wagoner, 1985; Van Wagoner et al., 1987, 1988, 1990).

systems tract: Sequences are subdivided into systems tracts (lowstand-, transgressive-, highstand- / shelf margin systems tracts; Vail et al., 1991) (Fig. 4-1). Systems tracts consist of one or several parasequence-set(s) and a set of depositional systems. Interpretation of systems tracts is based on parasequence stacking patterns, position within a sequence and kind of bounding surfaces. The definition of systems tracts is not directly related to the sea level development (Brown & Fisher, 1977).

depositional system: Three-dimensional lithofacies body with sediments deposited under similar facies conditions (comparable to ‘facies type’) (Vail et al., 1991: 621).

cycle: Series of events with return to starting point (Bates & Jackson, 1987).

accommodation space: Potential depositional space for sediments, which is created by rising sea level, subsidence or a combination of both processes.

Hierarchy of cycles: Sequences are classified into different order based on their duration (the following mainly after Emery & Myers, 1996):

- **First order** cycles have periods >50 My and attributed to changes in ocean volume related to plate tectonic cycles.
- **Second order** cycles (3-50 My) are commonly associated to changes in the rate of tectonic subsidence or rate of uplift in the sediment source terrane.
Fig. 4-1. Sequence stratigraphic concepts (from Haq et al., 1987). Depositional model showing systems tracts. A. Systems tracts in relation to depth. B. Systems tracts in relation to geologic time.

SURFACES
- SB1 = SEQUENCE BOUNDARIES
- SB 1 = TYPE 1
- SB 2 = TYPE 2
- IDLS = DOWNLAP SURFACES
- tfs = top fan surface
- tls = top leveed channel surface
- ITI = TRANSgressive SURFACE
- First flooding surface above maximum regression

SYSTEMS TRACTS
- HST = HIGHSTAND SYSTEMS TRACT
- TST = TRANSGRESSIVE SYSTEMS TRACT
- LSW = LOWSTAND WEDGE SYSTEMS TRACT
- mvs = incised valley fill
- ft = fan channels
- lfs = fan lobes

Fig. 4-2. Relationships between sedimentation, relative sea-level and shoreline trajectories in regressive settings. A. Forced regression. B. Normal regression (from Helland-Hansen & Gjelberg, 1994).
Third order cycles (0.5-3 My) are considered to be controlled by eustasy, although tectonic mechanisms are possible.

Fourth order (0.1-0.5 My) 'parasequence' cycles represent individual shallowing upward facies cycles bounded by abrupt deepening. These may be related in part to autecyclic processes within the sedimentary system. Fourth order sequences are also termed as 'high frequency sequences' (HFS's) (Mitchum & Van Wagoner, 1991).

SYSTEMS TRACTS AND BOUNDING SURFACES (Fig. 4-1)

lowstand systems tract (LST) and shelf margin systems tract (SMST):
The basal systems tracts in a type-1 (LST) and a type-2 (SMST) sequence. These systems tracts are bounded at the base by a sequence boundary (sb): Discrete, extensive surface, which at least in part of the basin is developed as an unconformity (Mitchum et al., 1977). Unconformities of this kind are formed if the rate of sea level fall exceeds the rate of subsidence (discussion of the lateral geometry of sequence boundaries in Cartwright et al., 1993).

type-1 sequence boundary: Shelf exposure reaches beyond the offlap break (shelf break), e.g. caused by a strong and fast eustatic sea-level fall. The type-2 sequence boundary is characterized by an unconformity on the entire shelf. The overlying sequence is termed a „type-1 sequence“ (Van Wagoner et al., 1988).

type-2 sequence boundary: Only part of the shelf falls dry (e.g. if rate of subsidence is greater than eustatic sea level fall). On the oceanward part of the shelf, the sequence boundary is characterized by a conformity (Van Wagoner et al., 1988). The overlying sequence is termed a „type-2 sequence“ which starts with a shelf margin systems tract (SMST). The Exxon group today interprets this kind of bounding surface only as a boundary of a parasequence-set and rejects the model of type-2 sequences.

The typical LST consists of three components (Emery & Myers, 1996): The initial*
  * basin floor fan unit is detached from the foot of the slope, the subsequent
  * slope fan unit abuts the slope and the
  * lowstand prograding wedge forms a topset / clinoform system.

forced regressive wedge systems tract: Set of downstepping prograding wedges which is deposited during falling relative sea level at ramp-type basin margins (in contrast to steeper basin margins which are characterized by sediment bypass) (Posamentier et al., 1992). Shoreline trajectory is dipping basinwards and is regressive (Helland-Hansen & Martinsen, 1996) (Fig. 4-2).

Sediments of the LST or SMST may be absent on the proximal shelf because these systems tracts are often deposited at the shelf slope (Fig. 4-1).

transgressive systems tract (TST):
The TST is the middle systems tract of both type-1 and type-2 sequences. It is deposited during that part of a relative sea level rise cycle when topset accommodation volume is increasing faster than the rate of sediment supply (Emery & Myers, 1996). The transgressive surface (ts) forms the basal boundary of the TST and is characterized by abrupt deepening (a flooding surface).

highstand systems tract (HST):
The HST is the youngest systems tract in either a type-1 or a type-2 sequence. It represents the progradational topset-clinoform system deposited after maximum transgression before a sequence boundary, when the rate of creation of accommodation is less than the rate of sediment supply (Emery & Myers, 1996). The HST is separated from the TST by a maximum flooding surface (mfs) which marks the beginning of progradation (downlap). Often development of a condensed section (Vail et al., 1991). Maximum landward extension of deep water facies into shallow water facies (Read, 1995: 301-32).

OTHER IMPORTANT ARCHITECTURAL ELEMENTS
flooding surface: Boundary between older and younger sediments with evidence of abrupt deepening (e.g. a transgressive surface)
ravinement surface: Erosional surface formed during transgression in the high energy coastal environment (Demarest & Kraft, 1987). In most cases diachronous due to restriction to the coastal facies. Ravinement surfaces often only represent parasequence boundaries. More significant ravinement surfaces form transgressive surfaces (Emery & Myers, 1996).
transgressive lag: Coarse grained sediment which has been reworked from underlying strata in the coastal zone during transgressions (thickness often less than 60 cm) (Van Wagoner et al., 1990).

condensed section (sensu stricto): Thin (hemi-) pelagites which are deposited during periods of landward retrogradation of parasequences and hence minimum terrigenous input (Loutit et al., 1988; Kidwell, 1991). Condensed sections in the broader meaning are all successions which are relatively thinner than any age-equivalent reference sections (Gómez & Fernández-López, 1994; see also for discussion of sequence stratigraphic applicability of condensed sections).

topset: proximal portion of the basin-margin profile characterized by low gradients (<0.1°). Topsets effectively appear flat on seismic data and generally contain alluvial, deltaic and shallow marine depositional systems (Emery & Myers, 1996) (Fig. 4-3).

clinoform: more steeply dipping portion of the basin margin profile (commonly >1°) developed basinward of the topsets. Clinoforms generally contain deeper water depositional systems characteristic of the slope (Emery & Myers, 1996) (Fig. 4-3).

bottomset: portion of the basin-margin profile at the base of the clinoform characterized by low gradients and containing deep-water depositional systems (Emery & Myers, 1996) (Fig. 4-3)

offlap-break: main break in slope in the depositional profile between topsets and clinoform (Vail et al., 1991) (Fig. 4-3)

STACKING PATTERN OF PARASEQUENCES (Fig. 4-4)
retrogradation / backstepping: Landward migration of parasequences (transgression), sedimentation rate smaller than rise of relative sea level.

progradation: Seaward migration of parasequences (regression), sedimentation rate greater than rise of relative sea level

aggradation: Parasequences stacked above one point without lateral shift, sedimentation rate corresponds to rise of relative sea level.

VERTICAL SEA LEVEL FLUCTUATIONS (after Posamentier & James, 1993) (Fig. 4-5)
eustatic sea level: Distance sea level / fixed point within the geoid (e.g. centre of the earth)
relative sea level: Distance sea level / fixed point within the sediment (Mitchum, 1977)
paleodepth: Distance sea level / sediment surface (=sediment-water boundary)

LATERAL COAST LINE MIGRATIONS (may be independent of vertical sea level fluctuations!)
transgression: landward shift of coast line

regression: seaward shift of coast line

normal regression: rising or constant relative sea level; sediment input greater than accommodation space (Posamentier et al., 1992).
forced regression: falling of relative sea level; regression independent of sedimentary input (Posamentier et al., 1992).

Fig. 4-3. Typical profile of a prograding basin-margin unit, comprising topsets and clinoforms separated by a break in slope, the offlap break. Bottomsets may also be present (from Emery & Myers, 1996).
Fig. 4.4. Depositional architecture as a function of accommodation volume and sediment supply (from Emery & Myers, 1996; after Galloway, 1989).

Fig. 4.5. Definitions of sea-level (from Emery & Myers, 1996; after Jervey, 1988).
4.2. Paleoecologic interpretations based on foraminifera and calcareous nannofossils

Besides their great biostratigraphic value, foraminifera and calcareous nannofossils also hold information about paleo-water depth and ocean surface water productivity. In order to get a maximum of paleoecologic information from these two microfossil groups and to avoid misinterpretations it is important to evaluate the basic oceanographic setting of the studied period and area. In chapters 6 and 7 two examples from the Late Maastrichtian and the Paleocene of eastern Sinai are presented where similar fluctuations in the planktonic-benthonic foraminiferal ratios are developed, however, control mechanisms and therefore the geologic interpretations differ markedly.

**Planktonic Benthonic Foraminiferal Ratio and Benthonic Foraminifera**

Productivity of planktonic and benthonic foraminifera is influenced by numerous water mass properties such as temperature, pressure, density, nutrients, salinity, light penetration, oxygen, as well as other physical, chemical and biological factors. The two main important processes capable of affecting these parameters are sea-level changes and climatic / oceanographic fluctuations. If the water mass properties are mainly controlled by water depth, physiobiochemical parameters and therefore the distribution patterns of planktonic and benthonic foraminifera often vary predictably along a depth gradient. One of the major changes represents the increase in the percentage of planktonic foraminifera in the bottom assemblages at greater depths. This reflects the trend that planktonic species occur in greater densities in open oceanic environments, whereas the abundance of benthonic foraminifera is higher in neritic environments (Phleger, 1964; Reiss et al., 1974; Gibson, 1989; Van der Zwaan et al., 1990). An increase in the foraminiferal planktonic benthonic (P/B) ratio may therefore be interpreted in a suitable environment as a period of deepening. Abundance of certain benthonic foraminiferal species can also be linked to paleodepth and may therefore be used for approximate paleobathymetric trend analyses (Murray, 1991; Speijer, 1994a,b). Previous studies have shown that many species and genera of benthonic foraminifera can be attributed to distinct broad depth ranges (Berggren & Aubert, 1975; Aubert & Berggren, 1976; Van Morkhoven et al., 1986; Saint-Marc, 1992; Speijer, 1994a; Luger, 1985) (Tab. 6-1).

If the water mass properties are dominated by climatically induced productivity changes (e.g. upwelling or sinking of saline, oxygen-poor, nutrient-rich water plumes), different trends in foraminiferal distribution are developed (e.g. Altenbach & Sarntheim, 1989; Widmark, 1995). In this case, the P/B ratio may be sensitive for productivity changes. A good example is provided by Leary et al. (1989) and Leary & Hart (1992) who recorded P/B changes of up to 50% across a paleoclimatically controlled chalk-marl couplet of the Cenomanian of Southern England. They demonstrated that the highest planktonic values were reached in chalks which they interpreted as having been deposited during climatically induced surface-water productivity blooms. Productivity fluctuations in the surface waters may also lead to significant changes in the benthonic foraminifera fauna. Leary et al. (1989) for example postulate that substrate changes were caused by climatically-induced surface water productivity blooms, which in turn may lead to a reorganisation of the benthonic foraminiferal assemblages. High productivity events may also cause oxygen depletion with a bloom of anaerobic forms (Kaiho, 1991). Detailed discussions on the distribution of planktonic and benthonic foraminifera are found in Be (1977), Boltovskoy & Wright (1976), Murray (1973, 1991), Vincent & Berger (1981), Gibson (1989) and Van der Zwaan et al. (1990).

**Calcareous Nannofossils**

Calcareous nannofossils are mainly used in biostratigraphy rather than as paleoenvironmental indicators due to their planktic mode of life and rapid distribution. Under certain conditions their distribution patterns may reflect the ecological conditions (e.g. surface water paleoproductivity) within the surface waters (Murray, 1995, Eshet et al., 1994; Eshet & Almagi-Labin, 1996). Nevertheless, sea-level reconstructions by means of calcareous nannofossils (e.g. Reale, 1995) are complicated (Perch-Nielsen, 1985).
5. Sequence stratigraphy of the Upper Cretaceous of central east Sinai, Egypt

*S. Luning, †A. M. Marzouk, §A. M. Morsi and *J. Kuss

* University of Bremen, FB5 - Geosciences, PO Box 330440, 28334 Bremen, Germany
† Tanta University, Faculty of Science, Geology Dept., Tanta 31511, Egypt
§ Ain Shams University, Faculty of Science, Geology Dept., Abbassia, Cairo, Egypt

Abstract

Deposition of the Upper Cretaceous of central east Sinai has been controlled by a long-term transgressive phase and several higher order sea level fluctuations. The paper gives a first sequence stratigraphic interpretation for this interval in the region, based on detailed sedimentological, biostratigraphical and paleoecological investigations in 13 Turonian-Maastrichtian sections and a review of all data available from the literature. Six main facies zones have been differentiated, including coastal mud flats with tidal channels, gypsiferous sabkha plains, peritidal siliciclastics, peritidal carbonates, high- and low-energy carbonate inner shelf facies as well as microfossil-rich outer shelf pelites. Biostratigraphy is mainly based on planktonic foraminifera, calcareous nannofossils, ostracoda, and ammonites. The study is restricted to an area that had been tectonically rather quiet during the Late Cretaceous-Early Tertiary, lying south of the Syrian Arc intraplate foldbelt which experienced major uplifting during this period. Within the Turonian to Maastrichtian interval, six major sequence boundaries have been reconstructed. Cycle duration varies between 4 to 9 Ma, which indicates a cycle order being intermediate between 3rd and 2nd. Correlation with other sea level reconstructions from the region (Egypt, Israel, Jordan, Tunisia) points to a more or less synchronous regional sea level development. Comparison of the regional sequences with the ‘eustatic’ model of Haq et al. (1987) involves uncertainties; nevertheless, some of the sea level fluctuations recorded in Sinai may be related to worldwide eustatic sea level changes.

5.1. Introduction

The Late Cretaceous of Sinai and of the Negev Desert is controlled by a long-term transgressive phase and several higher order sea level fluctuations with depositional environments ranging from terrestrial to hemipelagic. The depositional history of Sinai and of the Negev Desert from this period was previously studied by several authors (e.g. Bartov et al., 1972; Flexer et al., 1986; Cherif et al., 1989a, b; Lewy, 1990; Kuss, 1992; Orabi, 1992, 1993; Ziko et al., 1993; Kora & Genedi, 1995). Nevertheless, a comprehensive sequence stratigraphic model for these successions has not been published. This study, therefore, aims at giving a first sequence stratigraphic interpretation for the Upper Cretaceous of central east Sinai on the basis of detailed sedimentological, biostratigraphical, and paleoecological investigations in 13 Turonian-Maastrichtian sections (Fig. 5-1). The available literature on this interval in central east and central west Sinai has been reviewed and included in the sequence model. The study area is situated in a region that was tectonically rather quiet during the Late Cretaceous-Early Tertiary, lying south of the Syrian Arc intraplate foldbelt which experienced major uplift during this period. The sea-level changes described in this study are, therefore, expected to possess at least regional character and may be largely independent of local compressional processes in the foldbelt. However, larger-scale loading and relaxation effects in terms of a foreland model cannot be fully excluded.

5.2. Regional setting

During the Late Cretaceous, Sinai was part of the broad northern shelf of the Afroarabian Plate. The shelf margin is inferred to have been located near the present-day northern coastline of Sinai (Bein & Gvirtzman, 1977; Ginzburg & Gvirtzman, 1979). The Gulf of Suez and Gulf of Aqaba rifts, which bound the Sinai microplate today, were still closed. While northern Sinai and the adjacent Negev were affected by transpressive movements from the Late Turonian onwards, southern and central Sinai (the study area) are believed to have remained tectonically rather quiet throughout the Mesozoic and Early Tertiary (Said, 1962; Cohen et al., 1990; Kerdany & Cherif, 1990). The SE-vergent, NE-SW trending domal anticlines in northern Sinai is part of the “Syrian Arc” (Krenkel, 1924, 1925) which represents an intraplate foldbelt extending from Egypt to Syria and was formed by Late Cretaceous to Recent inversion of Late Triassic / Liassic half graben (Moustafa and Khalil, 1990, Chaimov...
Figure 5-1. Correlation of sequence boundaries in the studied sections. Stippled lines indicate uncertain position of the sequence boundary. Note that fill patterns indicate facies zones and not detailed lithologies. Biozones are abbreviated, planktonic foraminiferal zones in left column and calcareous nannofossils in right column. The datasets of sections B, M and T2 are presented in Figures 5-6 to 5-9. East-west running stippled line marks Themed Fault which separates the tectonically 'stable shelf' of central and southern Sinai from the 'unstable shelf' in northern Sinai (see Said, 1962).
et al., 1993, Shahar, 1994). Following the nomenclature of Said (1962), the inversion zone is also termed ‘unstable shelf’, and the southern tectonically calm block ‘stable shelf’ (Fig. 5-1). With one exception, the sections studied here are situated on the stable shelf.

Owing to the intermediate position between mainland Egypt and Israel, different lithostratigraphic systems have been used for the Upper Cretaceous in Sinai. While the Egyptian formations are mainly based on Ghorab (1961), the Israeli nomenclature was developed by Flexer (1968) and Bartov et al. (1972). Correlations of the different schemes (Fig. 5-2) are included, for example, in Kerdany & Cherif (1990), Kuss (1992) and Kora & Genedi (1995). In the following section, a short characterization of the different formations in the study is given, while detailed lithological descriptions and remarks on differentiation of the formations in central east Sinai can be found, for example, in Bartov et al. (1972), Ziko et al. (1993) and Kora & Genedi (1995). The Abu Qada Formation (=Ora Shales in the Israeli nomenclature) consists of mainly pelitic lithologies and is attributed to the late Cenomanian to early Turonian. These soft lithologies are overlain by hard limestones of the late Turonian Wata Formation (=Geroff Formation). The Coniacian-Santonian succession is included in the Matulla Formation (=Zihor Formation and lower part of Sayyarim Formation) consisting mainly of calcareous siliciclastics, carbonates and inner shelf shales. Towards the north, the siliciclastics are replaced by carbonates which are grouped into the Themed Formation in this area. These units are overlain by the Campanian-Maastrichtian Sudr Chalk (upper Sayyarim Formation and Ghareb Formation) which is composed of chalks, cherts and limestones. Isopach maps for the different Late Cretaceous formations can be found in Bartov & Steinitz (1977). It may be interesting to note that Upper Cretaceous deposits play a significant role in hydrocarbon plays in the petroliferous Gulf of Suez Basin (e.g. Sestini, 1995) as well as in the rare hydrocarbon discoveries in northern Sinai (see Alsharhan & Salah, 1996).

5.3. Materials and methods

During two field expeditions in 1995 and 1996, twelve Upper Cretaceous sections were measured in central east Sinai and one section at the Gebel Areif El Naqa anticline in northeast Sinai (Fig. 5-1). During fieldwork, special attention was paid to sedimentary structures and horizons with clear lithological changes. For thin section analysis, 48 handspecimens from limestones and sandstones were taken. For microfossil investigations, chalks, marls, and shales were sampled at intervals ranging between 1-3 metres with a total of 171 samples. For extraction of foraminifera and ostracoda the pelitic samples were washed twice with a 63 μm sieve after treatment with H2O2 and the highly concentrated tenside REWOQUAT, respectively. The microfossil residue was then fractionated into four grain size classes for easier handling. The planktonic foraminifera were identified under the light microscope (by Luning). For the calcareous nannoplankton, smear slides were prepared using techniques described in Bramlette & Sullivan (1961) and Hay (1961, 1965). The slides were examined (by Marzouk) under the light microscope at a magnification of about x1250 by both cross-polarized and phase-contrast. Ostracodes were picked from all samples and were investigated by Morsi.

5.4. Biostratigraphy

Biostratigraphy in this study is mainly based on planktonic foraminifera and calcareous nannofossils while macrofossils yielded only little supplementary data (see below). For palaeocologic reasons, biostratigraphy by planktonic foraminifera and calcareous nannofossils is restricted to the more hemipelagic parts of the succession. While the facies of the upper Coniacian, Santonian and Maastrichtian are often well suited, the shallow-marine deposits of the Turonian to middle Coniacian in the study area often do not contain biostratigraphically indicative species. Although intense silicification restricts microfossil preservation in several intervals of the mostly hemipelagic Campanian, planktonic foraminiferal faunas of varying abundance have been found in the unsilicified parts of the succession.

The Late Cretaceous hemipelagites of Sinai were previously biostratigraphically studied by means of planktonic foraminifera by several authors. While some workers focussed on specific time intervals (Cenomanian-Turonian: e.g. Cherif et al., 1989a; Orabi, 1992; Coniacian-Santonian: Ismail, 1993; Maastrichtian-Eocene: e.g. Said & Kenawy, 1956; Abdelmalik et al., 1978; Cherif et al., 1989b, Shahin, 1992), others elaborated on the whole Late Cretaceous interval (Shahin & Kora, 1991; Hewaidy et al., 1991; El Sheikh, 1995; Ayyad et al., 1996). In this contribution we use the general Tethyan biozonation scheme for planktonic foraminifera by Robaszynski et al. (1984), and for the Turonian to Lower Coniacian the scheme of Caron (1985). However, some regional
peculiarities exist, namely the absence of *Globotruncanita calcarata* in central east Sinai which is also supported by a study of Shahin & Kora (1991). This may in part be related to silicified-phosphoritic lithologies in the uppermost Campanian (Taba section, Figs. 5-7, 5-8) from which no planktonic foraminifera could be derived. Findings of *G. calcarata* were previously reported from northeastern Sinai (Hewaidy et al., 1991) and Israel (Almogi-Labin et al., 1986; Gvirtzman et al., 1989), while in other sections in Israel (e.g. Almogi-Labin et al., 1993) and western Sinai (Cherif et al., 1989b; Orabi, 1991) *G. calcarata* is missing. The occurrence of *G. aegyptiaca* already in the *G. calcarata* Zone was reported by Almogi-Labin et al. (1986) from Israel. From northeastern Sinai, *G. aegyptiaca* was described from the base of the *G. falsostaurti* Zone (Hewaidy et al., 1991). This contrasts the Tethyan scheme of Robaszynski et al. (1984) who postulate a first occurrence within the lower Maastrichtian representing the base of the *G. aegyptiaca* Zone of Caron (1985). Due to the general absence of *G. calcarata* and the early occurrence of *G. aegyptiaca*, the base of the *G. falsostaurti* Zone and therefore the Campanian-Maastrichtian boundary [see also general boundary discussion in Odin (1996)] remains rather poorly defined in the Taba section and is interpreted because the interval is directly overlain by deposits of the *G. gansseri* Zone and because of the the absence of *G. subspinosa*. Some characteristic planktonic foraminifera from the Upper Cretaceous of eastern Sinai are illustrated in Figure 5-3.

Biozonation schemes for benthonic foraminifera of the Upper Cretaceous in Sinai and the Gulf of Suez were presented by Ansary & Tewfik (1969), Shahin & Kora (1991), Hewaidy et al. (1991), and Hewaidy & El Ashwah (1993). However, because biostratigraphic resolution is very limited and occurrence of index species is considered to depend strongly on facies, benthonic foraminifera have not been biostratigraphically studied. Only a few studies of calcareous nanofossils have been published from the Late Cretaceous of Sinai (Arafa & El Ashwah, 1988; Arafa, 1991; El Sheikh, 1995). More extensive investigations have been carried out in neighboring Israel (e.g. Eshet & Moshkovitz, 1995). Biozonation of calcareous nanofossils in this contribution is based on concepts of Crux (1982) for the Turonian to early Campanian and of Eshet & Moshkovitz (1995) and Perch-Nielsen (1979) for the late Campanian to Maastrichtian. Correlation of calcareous nanofossil biozones with those of planktonic foraminifera shows varying degrees of correspondence with the schemes published by Bralower et al. (1995) for the Tethys, and Eshet & Moshkovitz (1995) for Israel. While for the late Coniacian to Campanian, a good correlation with these schemes has been observed, the Maastrichtian correlation comes closer to the schemes in Bolli et al. (1985: 5).

Main differences in the correlation of our data from the Taba section with the schemes in Bralower et al. (1995), Eshet & Moshkovitz (1995) and Gvirtzman et al. (1989) are the rather early first occurrences of *Lithophidites quadratus* near the base of and *Micula murus* in the upper *Gansserina gansseri* Zone. The differences may be explained at least in part by the strong paleoecologic susceptibility of the two species. While *L. quadratus* is interpreted to be characteristic of low productivity conditions, *M. murus* has been found to bloom in high productivity settings (Lüning et al., subm. a). Some characteristic calcareous nanofossils from the Upper Cretaceous of eastern Sinai are illustrated in Figure 5-4.

Detailed biozonation schemes based on ostracoda have been proposed for the Cenomanian-Turonian (Rosenfeld following pages:

**Figure 5-3.** Planktonic foraminifera from the Late Cretaceous of eastern Sinai (sections A5, A6, A7, C, D; sections A5 and A7 near A6), scale bars = 100 μm. 1-2: *Abathomphalus mayerorum* (Boll.), late Maastrichtian. specimen 1 from sample A7-11, specimen 2 from sample A7-15. 3: *Plummerina reicheli* (Brönnimann) [sensu Masters, 1993], latest Maastrichtian. both specimens from sample A5-17. 4: *Globotruncanita aegyptiaca* Nakkady, late Maastrichtian, specimen 4 from sample C1-5, specimen 5 from sample C1-6, specimen 6 from sample C1-7. 7-8: *Dicerotina asymmetrica* (Sigal), Santonian. specimen 7 from sample A6-14, specimen 8 from sample A6-8. 9: *Whiteinella archaeocretacea* Pessagno, Santonian. specimen A6-5. 10-11: *Dicarinella concavata* (Brönnimann), Coniacian-Santonian, both specimens from sample D1-3. 12-13: *Marginotruncana sintusosa* Porthault, Coniacian-Santonian, both specimens from sample D1-3.

North Sea coast. The plant fragments in the upper Campanian marl point to a vegetated swampy marsh facies during this period. The diagenetic redox potential, related to bioturbation, hydrological and climatic conditions leading to a complex Coniacian-Santonian (section M). Red-green layering and mottling reflect strong vertical and lateral changes in related to a pronounced arid climate during this period (section B) and, although only in a few layers, during the development of gypsum is restricted to a sabkha setting developed mainly during the lower Turonian which is finely laminated, mottled reddish-greenish soft siltstones, shales (Fig. 5-10A) and (only in the lower Turonian) laminitation, ripples and silt-shale alternations reflect episodic flooding events with varying intensities. Intercalated fine sandstones (e.g. in the lower Turonian of section B and the Coniacian-Santonian of section M) represent deltaic and tidal distributary or crevasse channel deposits. The fine grained siliciclastic sediments are interpreted to have been deposited in silty/shaly intertidal mudflats located in deltaic and tidal interdistributary areas which have been sourced by overbank and crevassing processes. Lamination, ripples and silt-shale alternations reflect episodic flooding events with varying intensities. Intercalated fine sandstones (e.g. in the lower Turonian of section B and the Coniacian-Santonian of section M) represent deltaic and tidal distributary or crevasse channel deposits. The development of gypsum is restricted to a sabkha setting developed mainly during the lower Turonian which is related to a pronounced arid climate during this period (section B) and, although only in a few layers, during the Coniacian-Santonian (section M). Red-green layering and mottling reflect strong vertical and lateral changes in the diagenetic redox potential, related to bioturbation, hydrological and climatic conditions leading to a complex \( \text{Fe}^2+ / \text{Fe}^3+ \) distribution. Similar colour patterns can be found in modern tidal flats, for example at the German North Sea coast. The plant fragments in the upper Campanian marl point to a vegetated swampy marsh facies during this period.

5.5. Facies zones

In this study, six facies zones are differentiated in the Upper Cretaceous of central east Sinai. The main depositional elements include coastal mud flats with tidal channels, gypsiferous sabkha plains, peritidal siliciclastics, peritidal carbonates, high and low energy carbonate inner shelf facies as well as microfossil-rich outer shelf pelites. An idealized facies block diagram summarizing all facies zones developed during the Late Cretaceous in the study area is given in Figure 5-5. The diagram is intended as a reference for the interpretation of vertical facies changes in terms of Walther’s facies law and is used to illustrate potential positions of facies interfingering (see descriptions below). This facies model has to be regarded as a predictive tool and not as a paleogeographic reconstruction for a certain time interval, and thus no scale and no orientation are given. Detailed paleogeographic maps of central Sinai within a sequence stratigraphic framework, in turn, are presented and discussed further below. Concerning the interpretation of the facies diagram it has to be noted that, for a certain time slice, central east Sinai was dominated by only one to a maximum of three main facies zones (see below), meaning that the study area represents only a small part of a much larger continent-basin profile.

In the following sections, the different facies zones are described and an environmental interpretation is given. Previous facies investigations in central-east Sinai were carried out by Bartov et al. (1972), Lewy (1975), Orabi (1992, 1993), Ziko et al. (1993), Khalifa & Eid (1995), Kora & Genedi (1995), Ahmed (1995) and their results as well as data from neighbouring regions have been included in our model.

5.5.1. Supra- and intertidal mudflats, sabkha

Description. This facies is mainly developed in the Lower Turonian and Coniacian-Santonian as well as in short intervals in the Upper Turonian and Upper Campanian. The successions are characterized by alternating, partly finely laminated, mottled reddish-greenish soft siltstones, shales (Fig. 5-10A) and (only in the lower Turonian) primary gypsum. In many cases, the siltstones are crossbedded with occasional intercalations of channel fills of fine sandstone. In the Coniacian-Santonian of Taba (Fig. 5-7), reddish-brown clay horizons are developed. In the phosphoritic upper Campanian of the same section (Fig. 5-8), marl with abundant plant fragments (Fig. 5-10C) and devoid of foraminifera has been found.

Interpretation. The fine grained siliciclastic sediments are interpreted to have been deposited in silty/shaly supra- to intertidal mudflats located in deltaic and tidal interdistributary areas which have been sourced by overbank and crevassing processes. Lamination, ripples and silt-shale alternations reflect episodic flooding events with varying intensities. Intercalated fine sandstones (e.g. in the lower Turonian of section B and the Coniacian-Santonian of section M) represent deltaic and tidal distributary or crevasse channel deposits. The development of gypsum is restricted to a sabkha setting developed mainly during the lower Turonian which is related to a pronounced arid climate during this period (section B) and, although only in a few layers, during the Coniacian-Santonian (section M). Red-green layering and mottling reflect strong vertical and lateral changes in the diagenetic redox potential, related to bioturbation, hydrological and climatic conditions leading to a complex \( \text{Fe}^2+ / \text{Fe}^3+ \) distribution. Similar colour patterns can be found in modern tidal flats, for example at the German North Sea coast. The plant fragments in the upper Campanian marl point to a vegetated swampy marsh facies during this period.
Figure 5-5. Idealized facies block diagram summarizing all facies zones developed during the Late Cretaceous in central east Sinai. The figure points to positions of potential facies interfingering and serves as a reference for the interpretation of vertical facies changes according to Walther's facies law. Examples for typical facies transitions and vertical facies changes are given in the text. The main depositional elements in the Late Cretaceous of the study area include coastal mud flats with tidal channels, gypsiferous sabkha, peritidal siliciclastics, peritidal carbonates, high and low energy carbonate inner shelf facies and microfossil-rich outer shelf pelites. The diagram is meant as a generalized facies model to be used as a tool rather than to represent a paleogeographic reconstruction for a certain interval (for paleogeographic maps see Figure 5-15), therefore no scale and no orientation are given. Lithological legend as in Figure 5-8C. Using the diagram, it has to be kept in mind that quantitative distribution and geometry of the different facies zones may change significantly in relation to a number of parameters, such as sea level, paleoclimate and siliciclastic input.
Figure 5-6. Section B (N' Sheikh Attiya).
Legend in Figure 5-8C, location map in Figure 5-1.
Figure 5-7. Section M (Taba), lower part. Legend in Figure 5-8C, location map in Figure 5-1.
Figure 5.8. A Section M (Taba), upper part. B Enlargement of the interval 175-179 m from A. C Legend. Location map in Figure 5-1.

Legend for sections

- M. marginita
- M. schwageri
- M. sinali
- G. altifrons
- A. biowi
- A. cretacea
- H. muski
- A. parvus expansus
- Q. gartner
- L. quinquifolius
- S. cretulata
- R. angustus
- T. phaeolus
- R. anthophorus
- G. diplogrammus
- E. eximius
- Z. embergeri
- T. decorus
- W. bernessae
- K. magnificus
- L. olykura
- L. arcusius
- Q. trifidum
- Cytheris rosensis
- C. rosanfieldi
- C. rosanfieldi
- M. murus
- G. antipyracea
- G. ventricosa
- G. falkosturiati
- G. antillae
- G. conica
- G. stuartiformis
- G. balkei
- G. petenrei
- G. gansseri
- R. nigosa
- R. hexamerata
- R. subcircumstellar
- R. subpenetral
- E. eximius
- R. anthophorus
- P. cretacea
- A. cymbfornis
- M. staurophora
- M. decoratus
- P. stolven
- D. diplogrammatoides
- L. cay euxi
- L. quadratus
Figure 5-9. A Section T2 (between Taba and Nuweiba). B Section D (N’ Sheikh Attiya). Legend in Fig. 5-8C; location map in Fig. 5-1.
5.5.2. Siliciclastic delta and tidal sand ridges

Description. Quartzose fine to medium sandstones of this facies are developed in the Coniacian-Santonian and are intercalated in carbonates. The fine sandstones are composed of evenly fine laminated units alternating with dm-scale erosive crossbedded intervals (Figs. 5-12B-D) and sharply bounded, intensely slumped horizons. Slump fabric is characterized by asymmetric slump folds, slump-fault planes and slump-brecciation of cm-scale (Fig. 5-12B-D). In the Taba section (section M), pseudonodules of coarse to medium sandstone have sunk into layered fine siltstone and have deformed the underlying strata (Fig. 5-12F). This process is likely to occur, if a field of coarser-grained sand ripples migrates over unconsolidated finer-grained sediment which leads to loading effects and pseudonodule formation (Allen, 1982: Vol. 2: 360). The medium sandstones are arranged in tabular crossbeds of m-scale with vertically variable dip directions or are horizontally bedded. Locally, discrete Thalassinoides layers of cm-dm scale are developed. In the Taba section (section M), a clear coarsening upward trend from siltstone/fine sandstone to medium sandstone can be recognized within a 10 m interval (Fig. 5-7).

Grain composition reflects a high sandstone maturity. Grains consist predominantly of well to moderately sorted quartz and accessory feldspar and are angular to sub-angular / sub-rounded (Fig. 5-12G). All sandstones contain varying amounts of calcareous, often dolomitic matrix with lithological transitions to sandy (dolo-) wackestones, the latter characterized by the occurrence of bivalves, oysters, green algae, ostracodes, and gastropodes (Fig. 5-12I). The interfingering of siliciclastic and carbonate facies is also documented in fine alternations (mm-cm-scale) of rippled fine to medium sandstone and dolomitic micrite (Fig. 5-12H). In other examples, clusters of quartz grains accumulated in a dolomitic matrix. At some places in the Santonian and Campanian (e.g. in section T2), the sandstones possess a phosphoritic character with the phosphorite contained in the matrix or in the poorly sorted, angular to peloidal phosphoritic components which also include teeth and bone fragments. Often, glauconitic grains are found. Porosity in all sandstones is relatively high.

In central east Sinai, the Coniacian-Santonian siliciclastic interval was previously studied by Khalifa & Eid (1995). A comprehensive description of these siliciclastics from SW Sinai and the Gulf of Suez was published by Refaat (1993) and Enani et al. (1994) based on surface sections and offshore cores. Enani et al. (1994) convincingly showed that the sandstones in SW Sinai were deposited as ENE-oriented linear bar structures being several tens of km long, 5-10 km wide and 3-20 m thick.

Interpretation. The detailed studies of the Coniacian-Santonian siliciclastics by Enani et al. (1994) in western Sinai and the Gulf of Suez showed that the sandstones in that area are arranged in ENE oriented linear tidal bars with the long axis perpendicular to the coastline. By analogy, a similar interpretation is made for the Coniacian-Santonian sandstones in central east Sinai. However, orientations of the bars may be different because the coastline in central Sinai during this period was strongly curved. In the study area bars are inferred in two sections (sections M and D). In the Coniacian of the Taba section (section M), a typical shallowing upward trend, including the sandbar facies, is developed. The succession begins with a low energy inner carbonate shelf facies, followed by finely laminated, partly flasered or rippled siltstones and fine sandstones which are interpreted as deposits of the lower shoreface or offshore transition. This assumption is based on the vertical position of the unit within the succession, the typical sedimentary structures with lack of bioturbation pointing to reworking by (storm) waves, and the slumped horizons which suggest the presence of a gradient. The overlying tabular crossbedded medium sandstones are attributed to tidal sandstone bars sensu Enani et al. (1994) while the following crossbedded sandy oyster floatstones point to an inter-bar interval. The shallowing upward succession is topped by a typical reddish silty mud flat facies with dissecting small sandy channels.

Proximity and interfingering of the tidal sandstone facies with the nearby inner neritic carbonate facies is demonstrated by several features, namely the abundant calcareous matrix of the siliciclastics. Lithologic transitions to sandy bioclastic wackestones and small-scale sandstone-micrite alternations. While the high maturity of the sandstones with predominantly quartz grains points to considerable transport on the African Craton, the often subangular grain shapes seem to exclude very long transportation and indicate more regional source areas. The phosphoritic character of the Santonian-Campanian sandstones is connected with basinwide upwelling processes at the southern Tethyan margin during this period.

5.5.3. Calcareous peritidal

Description. In central east Sinai, the peritidal facies has only been found in the upper Turonian. A typical example is developed in the upper Turonian near Themed (section N1) where several subtidal-intertidal shallowing upward cycles of dm-scale were observed (Fig. 5-10B). While the subtidal lower parts of the cycles are composed of strongly bioturbated (Thalassinoides) wackestones and packstones containing miliolids, gastropodes, and oysters, the upper parts of the cycles are characterized by micritic laminites with even or wavy lamination. In some intervals, laminated stromatolitic mats have been found which, according to Monty (1967), are mainly developed in the intertidal zone. In other localities (e.g. section B), inversely graded peloidal layers with fenestral and birds-eye fabrics alternate with dark micritic horizons containing angular micritic clasts (Figs. 5-10E-F).
Figure 5-10.
Supratidal, calcareous peritidal and low energy carbonate inner shelf facies.

A. Post-CoSin transgressive surface separating inner shelf-type grainstones (lower TST) from underlying laminated reddish-greenish supratidal shales, siltstones and fine sandstones (upper LST); upper Coniacian, section M (Taba, ts at 69 m in Figure 5-7).

B. High frequency parasequences consisting of (laminated) tidalites and subtidal bioturbated horizons; Upper Turonian of section N1.

C. Supratidal marl with abundant plant fragments close to the Campanian-Maastrichtian boundary representing the post Ca/MaSin LST deposits in section M (Taba, at 177 m in Figure 5-8).

D. Upper Turonian peritidal stromatolite in section N1.

E-F. Fenestral fabrics (E) and inverse grading (F) in peritidal peloidal carbonates from the Upper Turonian of section B (at 184 m in Figure 5-6); photograph from thin section, sample B1-18.

G. Dolomitic wackestone with Acicularia (centre), serpels and molluscan shells; photograph from thin section, section G, sample G1-7.

H. Dolomitic micrite with oyster shells; photograph from thin section, section G, sample G1-7.

I. Dolomitic miliolid wacke-/packstone with reworked micritic clasts and chert concretions, transition between low and high energy calcareous inner shelf facies. photograph from thin section, section G, sample G1-5.
Figure 5-11.
High and low energy calcareous inner shelf facies.

A. *Thalassinoides* traces in the Turonian low energy carbonate inner shelf facies of section B (at 128 m in Figure 5-6).

B. Cross-stratified packstones with basal erosive flute casts cut into underlying calcareous marl (Upper Turonian of section B, at 122 m in Figure 5-6).

C. Wave ripples in oolitic packstone (Upper Turonian of section H).

D. Rip-up clast in cross-stratified dolomitic bioclastic packstone (Upper Turonian of section B, at 136 m in Figure 5-6).

E. Well sorted dolomitic packstone with *Acicularia magnapora* Kuss and oyster debris (section G, photograph from thin section G1-8).

F. Amalgamated high energy oolite with intercalated mud layers and relictic mud drapes in ripple troughs which had been deposited during low energy periods (Upper Turonian of section B, at 161 m in Figure 5-6, negative print from sample B1-15).

G. Oolitic grainstone with large reworked micritic clasts (Upper Turonian of section B, at 163 m in Figure 5-6; photograph from thin section B1-16).

H. Grainstone with abundant micritic clasts and *Dicyclina* sp. (Upper Turonian of section H, photograph from thin section H1-5).
Figure 5-12.
Chert and siliciclastic peritidal facies.

A. Autobrecciation in Campanian chert caused by diagenetic expansion (section M, at 160 m in Figure 5-8).

B-D. Slump horizon with slump folds (B) and slump shear planes intercalated in finely laminated undeformed fine sandstones (D); Coniacian, post-CoSin LST in section M (at 52 m in Figure 5-7).

E. Cross-bedded medium sandstones in the Coniacian of section M (at 55 m in Figure 5-7).

F. Coarse sandstone pseudonodules which have sunk into fine sandstones as an effect of loading (see Allen 1982: 360); Coniacian of section M (at 94 m in Figure 5-7).

G. Phosphoritic quartzose coarse sandstone, quartz grains subangular, dark grains phosphoritic; Coniacian-Santonian in section T2 (at 5 m in Figure 5-9A); photograph from thin section T2-3.

H. Fine-scale alternated bedding of dolomitic micrite and rippled fine to medium sandstone; Coniacian of section M (at 76 m in Figure 5-7); photograph from thin section M2-25.

I. Micrite with abundant angular quartz grains and molluscan shells; Coniacian of section M (at 71 m in Figure 5-7); photograph from thin section M2-23.
Figure 5-13.
Outcrops in Taba and macrofossils.

A. Significant flooding surface (fs) within the post TuSin-TST between Upper Turonian oolites and soft marly outer shelf deposits containing planktonic foraminifera. The sequence boundary lies in the uppermost part of the illustrated succession and is overlain by siliciclastics; section M (fs at 32 m in Figure 5-7).

B. Coniacian to Campanian succession in Taba section. While siliciclastics dominate the Coniacian, a more hemipelagic facies with cherts and (partly silicified) chalks prevails from the Santonian to Campanian. Corresponds to 130-160 meters in Figure 5-8A.

Ca-c and Da-b. *Coffaticeras* sp.; lower Turonian in section B (at 13 m in Figure 5-6).

E. *Solenoceras humei*. Campanian-Maastrichtian boundary just below the sequence boundary Ca/MaSin (section M, at 176.5 m in Figure 5-8A+B).

F+G. Porifera from the late Maastrichtian *Abathomphalus mayaroensis* Zone in section P.
Interpretation. The calcareous peritidal facies has only developed in settings with little terrigenous input within the studied interval. Consequently, suitable conditions were only developed during the late Turonian, because during the early Turonian and Coniacian/Santonian, the peritidal zone had been strongly influenced by siliciclastic mud flats and (only during the Coniacian/Santonian) deltaic tidal bars.

5.5.4. Calcareous inner shelf - low energy

Description. The calcareous, low energy inner shelf facies is mainly developed in the upper Turonian and, in the northern part of the study area, in the Coniacian as well. This facies consists of partly dolomitic wackestones, containing miliolids (Fig. 5-101) (e.g. Quinqueloculina sp.) and other benthonic foraminifera, ostracodes, gastropods, green algae (Fig. 5-10G), molluscan shells (including oysters, Fig. 5-10H), micritic clasts, peloids, echinites and bryozoans. Planktonic foraminifera are missing or very rare. Quantitative distribution of the different components varies widely in the studied successions. Observed end members include miliolid wackestones, oolitic wackestones, wackestones with green algae and molluscan shells, wackestones with oligospecific benthonic foraminifera and very few planktonic foraminifera, micritic oyster floatstones, gastropod wackestones and floatstones, echinite floatstones. Some of the low energy intervals are strongly bioturbated by Thalassinoides burrows (Fig. 5-11A). A Coniacian low energy oncoid belt in the Themed area with an N-E extension has been described by Lewy (1972).

Interpretation. See higher energy calcareous inner shelf facies.

5.5.5. Calcareous inner shelf - higher energy

Description. The higher energy inner shelf facies can be found mainly in the upper Turonian and in the northern part of the study area also during the Coniacian, comparable to the stratigraphic distribution of the low energy calcareous facies. The high energy facies is composed of well sorted packstones and grainstones with characteristic sedimentary structures such as dm-scale low-amplitude trough cross stratification, wave and current ripples (Figs. 5-11C+F), normal grading, erosive bases, flute casts (Fig. 5-11B) and sharply bounded conglomeratic tempestitic intervals (dm-m scale) which are often composed of dolomitic clasts. Composition and fabric of the components suggests that they had been reworked from the low energy facies (see above). The main components are green algae (Fig. 5-11F), gastropods, ostracods, echinite spines, peloids, and benthonic foraminifera (Fig. 5-11H). Often, the components are surrounded by one to a few micritic layers so that a superficial oolith fabric resulted (Fig. 5-11G). In some cases, intervals with multilayered ooids have been found. Typical microfacies types include oolitic grainstones with abundant green algae, oolites with molluscan shells and gastropods and green algae-gastropod grainstones. Further evidence of reworking is given by abundant, large, angular micritic clasts (Figs. 5-11D-G) and micritic internal sediment in molluscan tests which have been found in many cases. Porosity of the oolites is relatively high with cores leached in many cases. In the upper Turonian of section N1, several horizons of reworked rudists have been found which indicate the presence of rudist buildups.

In the Campanian and lower Maastrichtian of some localities, phosphatic higher energy deposits are developed, containing abundant subrounded, structureless phosphoritic clasts and fish teeth. These rocks have a dark brown to black weathered appearance in the field and possess a characteristic smell when crushed.

Interpretation. The calcareous inner shelf is interpreted to consist of a complex, patchy system of higher energy shoals which accumulated material reworked from low energy micritic inter-shoal areas. Such a relationship is suggested by the clear evidence of reworking and facies interfingering presented above. During redeposition in the shoals, the reworked components were often encrusted by thin superficial oolithic layers. Rather seldomly, this process continued to produce multilayered ooids, probably caused by restricted energetic conditions. Although statistical, high resolution microfacies analyses have not been carried out in central Sinai for the Late Cretaceous, the differentiation into several microfacies end members suggests that a number of subenvironments may be recognized. The low energy microfacies include lagoons dominated by miliolids, ostracods, green algae, gastropods, echinids, smooth-shelled oysters or combinations of these organisms. Composition of the shoals, however, may be strongly dependent on the type of the neighbouring low energy facies which provided most of the material for the shoals. Rudist banks and oyster bioherms (Bartov & Steinitz, 1982) represent higher energy subfacies with significant autochthonous carbonate production. Because the facies assemblage described is interpreted to possess a relatively high potential for autocyclic migration processes (see also Powell et al., in press), lateral reconstruction of carbonate subfacies as well as detailed sea level interpretations based on (non-peritidal) inner shelf carbonate facies distribution, is complicated and therefore was not attempted in this study.

5.5.6. Hemipelagic outer shelf

Description. The hemipelagic facies is widely developed from the Santonian to Maastrichtian of central east Sinai. Similar facies, although with a shallower character, have been found around the Cenomanian-Turonian
5.6. Sequence stratigraphy

Sequence reconstructions from outcrop-scale data require the completion of several steps during the course of data analysis (see e.g. Van Wagoner et al., 1988). In our study, the steps were as follows: 1) Facies interpretation for all sections was completed. 2) Horizons with facies changes were identified and vertical stacking patterns interpreted. 3) The cycles were then attributed to a 'high' and a 'low' order, based on their duration, amplitudes and stacking patterns. 4) Cycles from different sections were correlated within a biostratigraphic and partly lithostratigraphic framework.

Within the Turonian to Maastrichtian interval, six major sequence boundaries of 'low order' have been reconstructed (Fig. 5-14). Differentiation between type 1 and type 2 sequences (Van Wagoner et al., 1988) is complicated in this study because the investigated area of central Sinai represents only a small part of the continent-basin profile so that seismic-scale reconstruction of the whole profile cannot be achieved. Low order cycle duration varies between 4 to 9 Ma which indicates a cycle order being intermediate between 3rd and 2nd. According to the systems in Weber et al. (1995) and Emery & Myers (1996), maximum error ranges for the dating of our systems tracts boundaries are estimated as ± 1-2 My. The amplitudes shown in the relative sea level curve (Fig. 5-14) are based on the facies zones and unconformities as described below for the different systems tracts in central Sinai. Facies distribution maps have been compiled for the study area on systems tract-scale (Figs. 5-15A-M) based on our data and descriptions from the literature. Higher (4th-5th) order order cycles are discernable in central east Sinai (see below) but were not reconstructed on a basin-wide scale because this area (Steinitz, 1970) and were interpreted as a result of an increase in volume of the siliceous sediment during diagenesis. In other horizons the flint is partly brecciated (Fig. 5-12A) while the under- and overlying strata remained undeformed. This feature is assumed to be associated with autobrecciation processes which have also been caused by diagenetic flint expansion (Kolodny, 1969; Kolodny & Garrison, 1994).

**Interpretation.** The hemipelagic character of this facies is indicated by the presence of varying amounts of calcareous nannofossils and planktonic foraminifera. The foraminiferoid planktonic/benthonic ratio, which under certain conditions may serve as a paleobathymetric proxy (Phleger, 1964; Reiss et al., 1974; Gibson, 1989; Van der Zwaan et al., 1990), has been determined routinely for all foraminiferal samples. However, strong paleoproductivity changes in the southern Tethyan surface waters during this period (Almogi-Labin et al., 1993; Eshet & Almogi-Labin, 1996) seem to exclude high resolution sea level reconstructions based on foraminiferal distribution patterns in this setting (see also Lüning et al., subm. a).

The formation of flint is interpreted to be associated with marine high productivity conditions leading to blooming of diatoms and radiolaria (Moshkovitz et al., 1983; Reiss, 1988; Almogi-Labin et al., 1993; W. N. Krebs/Amoco, pers. comm. 1996). Intercalations of the flints in hemipelagic marls and the presence of radiolaria suggest a hemipelagic character for most of the flints. Nevertheless, there are some exceptions, for example, episodically intercalated flint beds in a Coniacian intertidal mixed siliciclastic-carbonate succession (Taba, section M).
### Boundary Ages, Epochs and Stages Based on Gradstein et al. (1995) (Cretaceous)

<table>
<thead>
<tr>
<th>Time (My)</th>
<th>Maastrichtian</th>
<th>Campanian</th>
<th>Coniacian-Santonian</th>
<th>Turonian</th>
</tr>
</thead>
<tbody>
<tr>
<td>65-70</td>
<td>late</td>
<td>early</td>
<td>late</td>
<td>late</td>
</tr>
<tr>
<td>70-75</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>75-80</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>80-85</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>85-90</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

#### Planktonic Foraminifera Biozones
- A. mayaroensis
- G. gansseri
- G. falsostuarti
- G. calcarata
- G. ventricosa
- D. asymetrica
- D. concavata
- D. primitiva
- M. sigali
- H. helvetica
- W. archaeocretacea

#### Calcareous Nannofossil Biozones
- M. mums
- L. quadratus
- A. cvmbiformis
- Q. trifidum
- R. anthophorus
- L. maleformis
- M. staurophora
- E. eximius

#### Sequence Chart for Central East Sinai

<table>
<thead>
<tr>
<th>Sequence Chart</th>
<th>Relative Sea Level for Central East Sinai</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mar/DaSin</td>
<td>? mfs</td>
</tr>
<tr>
<td>Ca/MaSin</td>
<td>ts</td>
</tr>
<tr>
<td>CaSin</td>
<td>ts</td>
</tr>
<tr>
<td>? mfs</td>
<td></td>
</tr>
<tr>
<td>CoSin</td>
<td>ts</td>
</tr>
<tr>
<td>? mfs</td>
<td></td>
</tr>
<tr>
<td>TuSin</td>
<td>ts</td>
</tr>
</tbody>
</table>

#### Figure 5-14
Regional sequence chart and relative sea level curve for central east Sinai. No numerical scale for sea level curve implied. Sequence boundary names consist of an abbreviation of the stage followed by the suffix ‘Sin’. Stage boundary ages are based on Gradstein et al. (1995), correlation with planktonic foraminiferal biozones on Bralower et al. (1995) and correlation of foraminiferal and nannofossil zones on own data as well as Bralower et al. (1995) and Eshet & Moskovitz (1995) (see text for discussion); ts = transgressive surface, mfs = maximum flooding surface.

### 5.6.1. TuSin

LST (Fig. 5-15B). In the Sheikh Attiya section (section B), sequence boundary TuSin is characterized by a marked upward facies change from low energy outer-inner shelf marls and carbonates to a thick silty mudflat facies with local sabkhaic gypsum formation. This significant facies break is preceded by two short-term shallowing events indicated by intercalation of peritidal siltstone horizons into the low energy calcareous shelf facies. The stacking pattern reflects prograding parasequences of the late pre-TuSin HST. Within this HST succession (facies map in Figure 5-15A), an outer shelf marl was found to contain, besides the dominating benthonic foraminifera, a few planktonics indicating a biostratigraphic age of *Whiteinella archaeocretacea* or *H. helvetica* Zone (Fig. 5-6). The dating is based on the presence of *W. archaeocretacea*, *Heterohelix moremani* and *Whiteinella baltica*. The same interval holds the ammonite *Choffaticeras* sp. (determination by Dr. P. Luger; Bremen) which indicates a maximum age of late early Turonian (Lewy & Raab, 1976). The sequence boundary must be only slightly younger and is estimated as late early Turonian (*H. helvetica* Zone).

A mid-Turonian high sea level within the *H. helvetica* Zone was also described from Israel by Honigstein et al. (1989) and may be correlated with the TST/HST prior to sequence boundary TuSin. From central-east Sinai, Orabi (1992, 1993) presented evidence of a hemipelagic character of the lower part of the Lower Turonian Abu
G  TST + HST post-Sa/CaSin early Campanian (G. elevate Zone)

H  LST post-CaSin late Campanian (G. ventricosa Zone)

I  TST post-CaSin late Campanian (G. calcarata Zone)

K  HST post-CaSin latest Campanian (G. calcarata Zone)

L  LST post-Ca/MaSin early Maastrichtian (G. fabostrati’ Zone)

M  TST + HST post-Ca/MaSin middle Maastrichtian (G. fabostrati’ Zone)
Ovocytheridea apiformis has been found which, according to Reyment (1960), is characteristic of the Coniacian-has been described from the late Turonian-Coniacian (Honigstein & Rosenfeld, 1985). In section B, the species Spinoleberis yatvatacnsis macro. The HST of sections B (Fig. 5-6) and H holds the ostracods Zone by calcareous nannofossils) which indicates an early Coniacian age (Fig. 5-7). A few specimen of Dicarinella sp. were also found in the ?highstand deposits of section H but the exact biozone remains unclear. The post-TuSin deposits in the study area are characterized by peritidal and supratidal silty mudflats in which deltaic or tidal fine sandstone channels are intercalated. Locally, gypsum formed in sabkha settings. Comparable nearshore to continental deposits are developed in neighbouring outcrops in the southern Negev Desert. A subtropical fresh water lake facies with an abundant angiosperm flora has recently been described by Debruska (1997) from the upper member of the Lower Turonian Ora Shales from that region. Abdel Gawad et al. (1992) subdivided the Turonian at Gebel Nezzazat in central west Sinai into three lithofacies units. While we interpret their basal 'carbonate part' as TST or HST, the middle 'clastic part' may be underlain by sequence boundary TuSin and consequently represents the LST. Their uppermost 'carbonate part' reflects post-TuSin TST and HST deposits and is most probably separated from the middle 'clastic part' by a transgressive surface. The Turonian strata in their section is overlain by Coniacian sandstones (Abdel Gawad et al., 1992) with a basal contact interpreted as sequence boundary CoSin. In a study from the southern Negev / central east Sinai area, Bartov et al. (1972) [also included in Bartov & Steinitz (1977)] subdivided the Lower Turonian Ora Shales into three members. The lower member is reported to consist of shales and limestones with ammonites and is interpreted by us as pre-TuSin TST. According to Bartov et al. (1972), the middle member is mainly composed of dolomites, oolites, chert, marl, sand and sandy limestones and, in our opinion, represents post-TuSin HST. The upper member of the Ora Shales varies in composition from shaly marls to gypsum and sandstones. We interpret this unit as the post-TuSin LST. Bartov et al. (1972) reconstructed re-flooding with re-establishment of open marine conditions for the late Turonian with deposition of the Gerofit Formation (post-TuSin TST+HST). Finally, the authors interpret a shallowing at the beginning of the Coniacian Zihor Formation which is composed of mixed carbonate-siliciclastics (post-CoSin LST).

TST (Fig. 5-15C). The transgressive surface (ts) in the Sheikh Attiya section (B) is characterized by a facies change from sabkha sedimentation with gypsum-shale interbeds to inner shelf-type green shale with oysters. The ts is estimated to be of middle Turonian age (H. helvetica or M. sigali Zone) based on the stratigraphic context. Planktonic foraminifera have not been found around this horizon. However, the TST in sections B and M yielded ostracods of the species Neocyprideis vandenboldi (Figs. 5-6, 5-7) which represents a marker for the Turonian (Gerry & Rosenfeld, 1973). The same horizon in section M holds rare occurrences of Spinoberis yatvataeensis which, according to Rosenfeld & Raab (1974), is also characteristic of the Turonian. The post-TuSin TST is developed in three of the studied sections (B, M, N1) and is characterized by predominantly inner shelf carbonates of the low and higher energy type. The TST interval is mainly represented by the sediments of the Wata Formation (Gerofit Formation in Isreal). The presence of higher order sequences is marked by repeated short-term switches to peritidal conditions, for example, by silty marls with characean algae, cross-bedded fine sandstones and calcareous laminites with birds-eyes. The high frequency shallowing upward cycles in section N1 (Fig. 5-10B) are interpreted as high frequency parasequences. In the upper part of the TST higher order sequences are indicated by repeated occurrences of planktonic foraminifera which may be related to erosion at the sequence boundary TuSin. Although we cannot agree with some of their sea level interpretations, their lithological and faunal descriptions provide valuable data to illustrate the complex lateral facies relationships during the Turonian of Sinai, and in general support our sequence stratigraphic interpretation. To the east, a late early Turonian omission surface was reported from the Maktesh Ramon anticline of the Negev Foldbelt by Lewy & Avni (1988). The authors assumed that the origin of this omission surface is associated with 'local weak differential movements (Gartunkel, 1964)' which are 'superimposed on the regional shallowing with local exposure observed in central Israel (Buchbinder et al., 1983; Sandler & Zilberman, 1985)'. The corresponding ages suggest that the 'regional shallowing' mentioned may also be associated with the regional sea level fall at sequence boundary TuSin.

HST (Fig. 5-15D). The maximum flooding surface (mfs) in the Tabat section (M) is drawn between the late LST hemipelagic chalky marls (see above) and overlying harder calcareous marls which are probably of inner neritic origin (Figs. 5-7, 5-13B). Further to the south in the Sheikh Attiya section (B), the mfs is represented by the boundary between a chalky interval, containing some planktonic foraminifera, and overlying predominantly low energy inner shelf carbonates (Fig. 5-6). In section M, the late TST deposits fall within the D. primitiva Zone (E. eximius Zone by calcareous nannofossils) which indicates an early Coniacian age (Fig. 5-7). A few specimen of Dicarinella sp. were also found in the highstand deposits of section H but the exact biozone remains unclear. The HST of sections B (Fig. 5-6) and H holds the ostracods Spinoberis yatvataeensis macro which previously has been described from the late Turonian-Coniacian (Honigstein & Rosenfeld, 1985). In section B, the species Ovocytheridea apiformis has been found which, according to Reyment (1960), is characteristic of the Coniacian-
Santonian (Fig. 5-6). Within the uppermost HST of section H, abundant occurrences of *Neocyprideis vandenboldi* were recorded which so far have only been described from the Turonian (Gerry & Rosenfeld, 1973). However, this species seems to be also present in the Lower Coniacian if our correlations are correct. The slight discrepancy may be also related to the problems of defining the Turonian-Coniacian boundary by different fossil groups such as ammonites and foraminifera. Summed up, the biostratigraphic information provided by ostracods support the concept of an early Coniacian age for the post-TuSin HST deposits which in the study area are characterized by carbonates and partly marls of the low and higher energy inner shelf facies. While the HST in section M is relatively thin, an increased thickness is documented in section B.

### 5.6.2. CoSin

LST (Fig. 5-15E). In section M (Fig. 5-7) the sequence boundary CoSin is characterized by a facies change from partly phosphoritic inner shelf carbonates to a peritidal siliciclastic system. Further to the north and to the west (section H, Fig. 5-1) the boundary is interpreted to be located between chalky HST deposits with a few planktonic foraminifera overlain by cross-bedded inner shelf carbonates characterized by shallow water indicators such as ooids, ostracods and milliolid foraminifera. The sequence boundary CoSin is of early Coniacian age (*D. primitiva* or *D. concavata Zone* by planktonic foraminifera / *E. eximius* or *M. staurophora Zone* by calcareous nannofossils) based on the datings of the late pre-CoSin TST (see above) and from a higher order transgressive phase within the post-CoSin LST (*D. concavata / L. maleformis*, section D; Fig. 5-9B). A roughly age-equivalent sea-level fall has also been described by Orabi (1991) from western Sinai.

The LST in the Taba section (section M, Fig. 5-7) is characterized by a subtidal to supratidal siliciclastic shallowing upward succession (see description in above) dominated by sandstone bars, peritidal deltaic silty mudflats, and subordinate by sabkhaic bedded gypsum. Higher-order sequences are indicated by repeated intercalations of shallow marine carbonates (oysters packstones, oyster calcareous marls in section M and inner-outershelf chalks in section D; Fig. 5-9B) into a peritidal silty mudflat and sandstone bar facies. Relatively high amounts of planktonic foraminifera in the chalks of section D are interpreted to reflect a retrogradational stacking pattern during the latest LST. Downslope, towards the northern part of the study area, the siliciclastics are gradually replaced by inner shelf carbonates (e.g. in sections H, G, N2/N3). The lowstand deposits of section H are composed of inner shelf carbonates containing the ostracod *Bythocypris windhami*. Until now, this species is known from the Coniacian to Campanian (Butler & Jones, 1957), providing further evidence for a Coniacian age for this LST. In the middle of section G a green algal flora including *Acicularia magnaporus* Kuss (Fig. 5-11E, previously described from the middle Turonian of Abu Roash; Kuss, 1994) *Halimeda* sp. and *Neomeris cretacea* was found. A description of higher energy oyster bioherms within a low energy inner shelf carbonate facies, has been given by Bartov & Steinitz (1982) from the Negev Desert and Sinai, including Themed and the Sheik Attiya area. A marginal example of this facies is probably developed in section N2/N3. According to Bartov & Steinitz (1982), the oyster bioherms had been bound to submarine morphotectonic paleohighs and locally interfinger with sandstones.

In this study, higher order sequences within the post CoSin LST are developed in section G, where siliciclastically influenced carbonates and beds containing a few planktonic foraminifera are intercalated in a calcareous inner shelf facies succession marking short-term periods of shallowing and deepening, respectively. This observation is consistent with results from Khalifa & Eid (1995) who carried out a detailed facies analysis of the Coniacian-Santonian lowstand deposits of central east Sinai and subdivided the succession into five shallowing-upward cycles (parasequences). A typical cycle is reported to consist of (from base to top) subtidal green glauconitic shales, carbonate packstones, phosphatic dolostones and intertidal quartz-arenites. However, correlation of our parasequence pattern with the cycles of Khalifa & Eid (1995) is complicated, because those authors based their model on a lithostratigraphic rather than a biostratigraphic framework.

A detailed description of the late Turonian to early Santonian lithological and facies development for southern Israel and Sinai, including the interval from the pre-CoSin HST to post-CoSin TST, was given by Lewy (1975). His reconstructions are based on a high resolution biostratigraphic framework provided by ammonites. Our findings correspond well with many of his results and interpretations. Lewy (1975) described a shallowing around the Turonian/Coniacian boundary with a maximum regression during early late Coniacian times which we correlate with sequence boundary CoSin. Renewed flooding of the shelf was inferred by Lewy (1975), based on the deposition of latest Coniacian chalks. That interpretation corresponds well with the timing of the post-CoSin transgressive surface which is discussed below. However, Lewy mentions several tectonic events, including regional tilting and folding, which may apply mainly for the Syrian Arc region but, according to our observations, have no significant role on the stable shelf of southern and central Sinai. Upper Turonian and Coniacian strata are reported to be absent (Kassab & Ismael, 1994) in the Abu Zuneima area (western Sinai). This stratal gap may be due to erosion and, or, non-deposition at this locality during the Coniacian LST.

The north-south trend of the Coniacian paleoshoreline in central-west Sinai, as shown in Fig. 5-15E, is clearly visible in isopach and sand/shale ratio maps of the Lower Senonian (Refaat, 1993) and is also shown in the Coniacian paleogeographic maps of Said (1990: p. 446). A Late Turonian–Early Coniacian phase of siliciclastic facies progradation has been reconstructed for southern Egypt ('facies 3' in Van Houten et al., 1984) and may represent the succession from the post-TuSin LST to the post-CoSin LST. Investigations of the geochemistry,
microfacies and facies of the mixed siliciclastic-carbonate LST sediments of the ‘Matulla Formation’ were carried out by Refaat (1993) in the Gulf of Suez region and Orabi & Ramadan (1995) in central west Sinai.

**TST and HST (Fig. 5-15F).** The TST and HST are composed of cherts and chalks, the latter usually containing significant amounts of planktonic foraminifera. Upper parts of the sequence (Santonian) are commonly phosphoritic. In general terms the interval corresponds to the middle and upper non-siliciclastic part of the Sayyarim Formation in Israel. Similar age-equivalent hemipelagites with a rich planktonic foraminifera fauna are described by Ismail (1993) from central-west Sinai. This interval has been described from Israel by Schneidermann (1970) and others.

The post-CoSin transgressive surface, interpreted in sections M and T2 (Figs. 5-7, 5-9A, 5-10A), separates peritidal siliciclastics to inner shelf carbonates from overlying, probably outer shelf cherty chalks and cherts. The age of the ts is estimated to be D. concavata Zone by foraminifera and as L. maleformis or R. anthropophorus Zone by calcareous nannofossils. This is based on datings of a higher order transgressive phase within the latest post-CoSin LST (D. concavata / L. maleformis Zones in section D, see above) and of the middle post-CoSin TST (D. concavata / R. anthropophorus Zones in section M). In the Nakhl-I well (data generously provided by the Gulf of Suez Petroleum Company, Cairo) Santonian hemipelagites of the lower D. asymmetrica Zone are recorded above sandy limestones, coinciding with a sharp drop in the resistivity curve. The age matches the data derived from our surface sections. It is interesting to note that the occurrence of the Santonian (Honigstein, 1984) ostracod Cythereis rosenfeldi rosenfeldi in the TST/HST of section M conforms with the dates provided by planktonic foraminifera and calcareous nannofossils (Fig. 5-8).

The exact position of the mfs is unclear. Either the HST is lithologically similar to the TST or was eroded during the subsequent sea level fall. From sections M (Fig. 5-8) and A6 the inferred mfs is estimated to have a biostratigraphic age of D. asymmetrica / L. cayeuxii.

5.6.3. Sa/CaSin

**LST.** In the study area, the sequence boundary Sa/CaSin has been found to be well developed in three sections. In a section south of Gebel Areif El Naqa (A6) in NE Sinai (Fig. 5-1), the boundary is macroscopically clearly marked by a pebbly intra-chalk unconformity with a probable hiatus. The section already lies on the unstable shelf. Because of its position in a synclinal area, deposition was less influenced by the Syrian Arc compression than in the anticlinal areas. In section M (Fig. 5-8), a hiatus even larger than that in section A6 is developed. Sequence boundaries Sa/CaSin and CaSin (see above) are amalgamated here, as demonstrated by a large biostratigraphic hiatus recorded by calcareous nannofossils. In section T2 (Fig. 5-9A) the sequence boundary is interpreted to lie between a partly silicified chalk and an overlying hard calcareous marl with oysters. The calcareous marl contains phosphatic pebbles and flint fragments at the base and may represent the lowstand deposits. In sections A6, M and T2 the sb is dated as lying close to the Santonian-Campanian boundary (D. asymmetrica-G. elevata boundary by foraminifera / L. cayeuxii-B. parca boundary by nannofossils). An age-equivalent sea level drop, marked by a clay bed intercalated in chalk throughout Israel, was also described (Lewy, 1990: 629). Based on the absence of a clear shallow facies, it is supposed that the post-Sa/CaSin LST was not deposited or has been eroded. From the Eastern Desert, Kuss & Malchus (1989) reported the absence of Santonian-lower Campanian which may be related to this sea level fall.

**TST + HST (Fig. 5-15G).** The transgressive surface in section T2 (Fig. 5-9A) is interpreted to lie at the lithologic boundary between a hard calcareous marl with oysters (see above) and an overlying softer chalky calcareous marl with planktonic-benthonic foraminiferal ratios ranging between 1-40% planktonic foraminifera. Because the LST is missing in section A6, the transgressive surface coincides with the sequence boundary. The hemipelagic sediments just above the Sa/CaSin unconformity in section A6, therefore, are interpreted as representing the lower TST. The ts here has an age of B. parca Zone (G. elevata Zone by foraminifera). The TST and HST are composed of chalk. The position of the mfs remains unclear in the studied sections and is estimated to be within the G. ventricosa Zone.

5.6.4. CaSin

**LST (Fig. 5-15H).** In section M, as described above, the sequence boundary CaSin seems to be amalgamated with sequence boundary Ca/SaSin. Here, the boundary is characterized by an erosive lower contact between hard calcareous marls and overlying chalky marls with a significant biostratigraphic hiatus, as demonstrated by the absence of two calcareous nannofossil biozones. The upper unit is interpreted as TST so that the LST in this section is missing. The biostratigraphic ages of the post-CaSin transgressive systems tract (Q. trifidum Zone in section M, Fig. 5-8) suggest that CaSin lies within these biozones or is older. A bored and abraded omission surface in chalk on paleohighs in Israel with comparable age has been described by Lewy (1990: 629) and may be related to the inferred sea-level fall at CaSin. Moreover, a fresh water diagenesis in these cherts has been reported by Kolodny et al. (1980).
The sequence boundary Ma/DaSin in the study area was interpreted around the K/T early Paleocene (NP1) marls with high foraminiferal planktonic values. The overlying Paleocene marls are interpreted to represent the post-Ma/DaSin TST and HST with typical shallow marine LST deposits missing 5.6.6.

Ma/DaSin sequence context.

Nevertheless, a late Maastrichlian age for the maximum flooding surface is assumed on the basis of the this hemipelagic setting is complicated because of the monotonous lithological record. In addition, it seems that their abundance distribution during the Santonian lo Maastrichtian (see detailed discussion in Luning et al., subm. central west Sinai (Cherif et al., 1989b) where the transgressive surface separates siliciclastics of the Duwi Formation from overlying chalks with planktonic foraminifera. Identification of the maximum flooding surface in the Campanian-Maastrichtian boundary is mentioned by Reiss (1988). Detailed sedimentological investigations of the late Campanian-early Maastrichtian phosphorites of Egypt (excluding Sinai) were carried out by Glenn (1990). He proposed a sequence stratigraphic development for this period which is similar to our model. Small differences in timing of the sequence boundary around the Campanian-Maastrichtian boundary (latest Campanian interpreted by Glenn versus earliest Maastrichtian in our study) may be largely attributed to the uncertainties involved in dating by planktonic foraminifera and other fossil groups (see discussion above).

Ca/MaSin

The sequence boundary Ca/MaSin in section M is characterized by a drastic facies change from inner shelf bioclastic phosphorites to marshy grayish-red marls devoid of microfossils, but with abundant plant fragments (Fig. 5-10C) The sequence boundary is assumed to lie within the Q. trifidum Zone and around the G. calcarata-G. falsostuarti boundary. Just below the sequence boundary within the latest HST, the ammonite Solenoceras hunei (Douville, 1928) was found which is interpreted by Abdel-Gawad (1990) to indicate the latest Campanian. Hiatuses around the Campanian-Maastrichtian boundary are also described from Israel (Shiloni et al., 1988; Almogi-Labin et al., 1990a; Lewy, 1990), from western Sinai (Orabi, 1991), and from the Eastern Desert (Luger, 1988; Hendriks & Luger, 1987). For Israel a fresh water diagenesis around the Campanian-Maastrichtian boundary is mentioned by Reiss (1988).

Detailed sedimentological investigations of the late Campanian-early Maastrichtian phosphorites of Egypt including Sinai) were carried out by Glenn (1990). He proposed a sequence stratigraphic development for this period which is similar to our model. Small differences in timing of the sequence boundary around the Campanian-Maastrichtian boundary (latest Campanian interpreted by Glenn versus earliest Maastrichtian in our study) may be largely attributed to the uncertainties involved in dating by planktonic foraminifera and other fossil groups (see discussion above).

TST and HST (Fig. 5-15M). The transgressive surface in section M is interpreted to lie between marshy marl deposits and overlying phosphoritic inner shelf carbonates. The transgressive surface is dated as Q. trifidum Zone / G. falsostuarti. While the lower TST is dominated by phosphoritic inner shelf carbonates, the upper part of the TST and the HST are composed of chalks and chalky calcareous marls containing abundant planktonic foraminifera and calcareous nanofossils (sections M, C, F, P). In section P, a hemipelagic sponge fauna occurs, including hexactinellids of the genera Rhizopoterion and Schicoradhus which were previously described from Maastrichtian-Paleocene chalks of the Egyptian Western Desert by Herrmann-Degen (1980). An early description of microfossils and fossil contents of the Maastrichtian white chalks of west Sinai was published by El-Shinnawi (1968).

A pronounced flooding event around the Campanian-Maastrichtian boundary is also described from sections in central west Sinai (Cherif et al., 1989b) where the transgressive surface separates siliciclastics of the Duwi Formation from overlying chalks with planktonic foraminifera. Identification of the maximum flooding surface in this hemipelagic setting is complicated because of the monotonous lithological record. In addition, it seems that foraminifera cannot be used for sea level reconstruction because of the strong influence of paleoproductivity on their abundance distribution during the Santonian to early Maastrichtian (e.g. Kolodny & Garrison, 1994). Phosphatic omission surfaces from the upper Campanian of southern Israel have been described by Soudry & Lewy (1990). They may be attributed to autocyclic changes in the energy regime rather than to major sea level falls.

5.6.5. Ca/MaSin

LST (Fig. 5-15L). The sequence boundary Ca/MaSin in section M is characterized by a drastic facies change from inner shelf bioclastic phosphorites to marshes grayish-red marls devoid of microfossils, but with abundant plant fragments (Fig. 5-10C) The sequence boundary is assumed to lie within the Q. trifidum Zone and around the G. calcarata-G. falsostuarti boundary. Just below the sequence boundary within the latest HST, the ammonite Solenoceras hunei (Douville, 1928) was found which is interpreted by Abdel-Gawad (1990) to indicate the latest Campanian. Hiatuses around the Campanian-Maastrichtian boundary are also described from Israel (Shiloni et al., 1988; Almogi-Labin et al., 1990a; Lewy, 1990), from western Sinai (Orabi, 1991), and from the Eastern Desert (Luger, 1988; Hendriks & Luger, 1987). For Israel a fresh water diagenesis around the Campanian-Maastrichtian boundary is mentioned by Reiss (1988).

Detailed sedimentological investigations of the late Campanian-early Maastrichtian phosphorites of Egypt (excluding Sinai) were carried out by Glenn (1990). He proposed a sequence stratigraphic development for this period which is similar to our model. Small differences in timing of the sequence boundary around the Campanian-Maastrichtian boundary (latest Campanian interpreted by Glenn versus earliest Maastrichtian in our study) may be largely attributed to the uncertainties involved in dating by planktonic foraminifera and other fossil groups (see discussion above).

TST and HST (Fig. 5-15M). The transgressive surface in section M is interpreted to lie between marshy marl deposits and overlying phosphoritic inner shelf carbonates. The transgressive surface is dated as Q. trifidum Zone / G. falsostuarti. While the lower TST is dominated by phosphoritic inner shelf carbonates, the upper part of the TST and the HST are composed of chalks and chalky calcareous marls containing abundant planktonic foraminifera and calcareous nanofossils (sections M, C, F, P). In section P, a hemipelagic sponge fauna occurs, including hexactinellids of the genera Rhizopoterion and Schicoradhus which were previously described from Maastrichtian-Paleocene chalks of the Egyptian Western Desert by Herrmann-Degen (1980). An early description of microfossils and fossil contents of the Maastrichtian white chalks of west Sinai was published by El-Shinnawi (1968).

A pronounced flooding event around the Campanian-Maastrichtian boundary is also described from sections in central west Sinai (Cherif et al., 1989b) where the transgressive surface separates siliciclastics of the Duwi Formation from overlying chalks with planktonic foraminifera. Identification of the maximum flooding surface in this hemipelagic setting is complicated because of the monotonous lithological record. In addition, it seems that foraminifera cannot be used for sea level reconstruction because of the strong influence of paleoproductivity on their abundance distribution during the Santonian to Maastrichtian (see detailed discussion in Luning et al., subm. a). Nevertheless, a late Maastrichtian age for the maximum flooding surface is assumed on the basis of the sequence context.

5.6.6. Ma/DaSin

Sequence boundary. The sequence boundary Ma/DaSin in the study area was interpreted around the K/T boundary and is marked in sections C and P by the presence of reworked Cretaceous planktonic foraminifera in early Paleocene (NP1) marls with high foraminiferal planktonic values. The overlying Paleocene marls are interpreted to represent the post-Ma/DaSin TST and HST with typical shallow marine LST deposits missing
(Luning et al., subm. b). In section M, the late Maastrichtian *A. mayaroensis* Zone is absent, which may be interpreted as a consequence of locally deep reaching erosion during the Ma/DaSin sea-level fall. Further evidence of the sea level drop around the K/T boundary comes from a number of other studies in the region. For example, very shallow conditions are reported from the K/T interval from central west Sinai by Ismail (1992) and Abbass et al. (1994) and from the Negev by Keller & Benjamini (1991) and Eshet et al. (1992). A hiatus at the K/T boundary is described from western Sinai from the Abu Zenima area (Anan 1992, Marzouk & Hussein, 1994, Gebel Makattah (Ayyad & Hamama, 1991), Wadi Feiran, Gebel Qabeliat (Marzouk & Abou-El-Enine, 1995) and several subsurface sections (Orabi, 1991). Reworked Cretaceous foraminifera have been observed during the *Pa. eugubina / Pa. pseudobulloides* Zones in Central Egypt (Luger et al., 1997, in press). The following Paleocene sea level development for central east Sinai has recently been described by Luning et al. (subm. b).

5.7. Comparison with sea-level reconstructions from neighbouring regions

5.7.1. Egypt, Israel, Jordan

The relative sea level curve proposed here for central east Sinai (Fig. 5-14) corresponds well to the reconstructions for Sinai and Israel by Reiss (1988) and Kuss & Bachmann (1996) (Fig. 5-16). A good correlation also exists with the cycle chart for Israel given by Flexer et al. (1986) except for the sea level drop associated with sequence boundary CaSin which is missing in their model (Fig. 5-16). Several significant mismatches were observed in the correlation of our reconstructions with the ‘Egyptian eustatic sea level’ chart by Said (1990: 449). A more complex sea level history compared to that observed in central east Sinai has been described by Lewy (1990) mainly from the unstable shelf of Israel (Fig. 5-16). Differences may be explained by tectonic movements of the Syrain Arc. However, the main cycles correlate well with our sea level model for central east Sinai. A sequence stratigraphic model for the Upper Cretaceous to Eocene of Jordan has recently been elaborated by Powell et al. (in press). Because their approach does not include biostratigraphic ages for their systems tracts, a direct correlation with our model is complicated. Nevertheless, their regional sea level reconstruction also seems to correspond largely to the situation encountered in central Sinai.

5.7.2. Arabian Peninsula

The Cretaceous sea level charts for the Arabian Peninsula (Harris et al., 1984) and the United Arab Emirates (Alsharhan & Kendall, 1995) in general differ from the sea level history in central Sinai because of strong tectonic activity in Arabia during that time. An exception is the K/T boundary for which those authors described the development of a pronounced unconformity.

5.7.3. Tunisia

A sequence stratigraphic study of the Turonian of west-central Tunisia has recently been presented by Saidi et al. (1997). They subdivided the Turonian into two sequences, one of late Cenomanian-middle Turonian, the other of middle to late Turonian age. The sequence boundary separating the two sequences is reported to lie within the *Helvetotruncana helvetica*, and the upper boundary of the second sequence is characterized by an early Coniacian period of shallowing (Saidi et al., 1997). Timing of these sequences corresponds well with the reconstructions for central-east Sinai and yields evidence of inter-regional processes controlling Turonian sedimentation in the southern Tethys. However, the facies in central-west Tunisia seems to be slightly deeper than in central east Sinai because the LST is described as inner shelf deposits and the TST/HST as containing significant pelagic elements (Saidi et al., 1997), contrasting with the peri- to supratidal post-TuSin LST and mainly inner shelf post-TuSin TST/HST in Sinai.

5.7.4. Eustatic sea-level curve

It is a standard procedure in every regional study to compare the reconstructed regional sea level cycles with the eustatic model published by Haq et al. (1987). However, today many stratigraphic intervals in the eustatic * Exxon* curve, especially the Late Cretaceous, are questioned in some respect (see Hancock, 1989, 1993; Miall, 1991, 1992). Because of apparent restrictions to the concept of a single eustatic curve it may be better in some cases to consider several regional curves, instead of one global curve. A comparison of the major sea level fluctuations of Sinai with those described in Haq et al. (1987) shows that the cycles in Sinai are in most cases longer than the 3rd order cycles in the ‘eustatic’ sea level chart. Considering the unavoidable uncertainties in sequence timing, for each sequence in Sinai several potential ‘eustatic’ sequences can be found so that convincing correlations can be seldom accomplished. This especially applies to the sequence boundaries in the lower Turonian (TuSin), around the Santonian-Campanian boundary (Sa/CaSin), and in the middle Campanian (CaSin) (Fig. 5-16). Higher probabilities for synchronism exist for the sequence boundaries of the lower Coniacian (CoSin), at the
Figure 5.16. Correlation of the relative sea level curve for central east Sinai (this study, position of sequence boundaries indicated) with other regional sea level reconstructions as well as the 'eustatic' model proposed by Haq et al. (1987). See text for discussion. For better comparison, the stage and biozonal scheme from Haq et al. (1987) has been kept, while a scheme which has been corrected for absolute ages and biozone correlation as realized in the study area, is presented in Figure 5.14. Note that the sea level reconstructions of Flexer et al. (1986), Reiss (1988), Lewy (1990), Said (1990), and Kuss & Bachmann (1996) in general were not developed within a high resolution biostratigraphic frame in terms of planktonic foraminifera and calcareous nannofossils so that correlation between the different curves involves some uncertainties.

<table>
<thead>
<tr>
<th>Stages</th>
<th>Planktonic Foraminifera biozones</th>
<th>Cacar nannofossil biozones</th>
<th>Regional transgressive-regressive cycles</th>
<th>Regional relative sea level curves</th>
<th>Data from Haq et al. (1987)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Maastrichtian</td>
<td>upper</td>
<td>NC22</td>
<td>G. gansser</td>
<td>NC21</td>
<td>1.1</td>
</tr>
<tr>
<td></td>
<td>lower</td>
<td>NC20</td>
<td>G. tricosatula</td>
<td>NC20</td>
<td>4.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>G. calcarata</td>
<td></td>
<td>4.4</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>G. ventricosa</td>
<td></td>
<td>4.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>G. elevata</td>
<td></td>
<td>4.2</td>
</tr>
<tr>
<td></td>
<td>upper</td>
<td>NC19</td>
<td>D. asymetrica</td>
<td>NC19</td>
<td>4.1</td>
</tr>
<tr>
<td></td>
<td>upper</td>
<td>NC18</td>
<td>D. concavata</td>
<td>NC18</td>
<td>3.6</td>
</tr>
<tr>
<td></td>
<td>upper</td>
<td>NC17</td>
<td>D. concavata</td>
<td>NC17</td>
<td>3.5</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>D. concavata</td>
<td></td>
<td>3.4</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>D. concavata</td>
<td></td>
<td>3.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>D. concavata</td>
<td></td>
<td>3.2</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>D. concavata</td>
<td></td>
<td>3.1</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>D. concavata</td>
<td></td>
<td>3.0</td>
</tr>
<tr>
<td></td>
<td>upper</td>
<td>NC15</td>
<td>D. primativa</td>
<td>NC15</td>
<td>2.9</td>
</tr>
<tr>
<td></td>
<td>middle</td>
<td>NC14</td>
<td>M. turgida</td>
<td>NC14</td>
<td>2.8</td>
</tr>
<tr>
<td></td>
<td>lower</td>
<td>NC13</td>
<td>H. helleriana</td>
<td>NC13</td>
<td>2.7</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>W. archaicaeulanae</td>
<td></td>
<td>2.6</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>R. custardani</td>
<td></td>
<td>2.5</td>
</tr>
</tbody>
</table>

Data from Haq et al. (1988)
Campanian-Maastrichtian boundary (Ca/MaSin) and at the K/T boundary (Ma/DaSin) (Fig. 5-16). A certain mismatch in the timing of CoSin may be related to differences in the lower limit of the *D. primitiva* Zone which in the chart of Haq et al. (1987) lies within the upper Turonian and in a scheme of Bralower et al. (1995) at the Turonian-Coniacian boundary.

In a 'eustatic' sea level reconstruction based on data from the tectonically stable Russian Platform and Siberia, Sahagian et al. (1996) described marked mid Turonian and mid Coniacian 'eustatic' sea-level falls which may be correlated with sequence boundaries TuSin and CoSin, respectively, in central east Sinai.

5.8. Conclusions

Deposition of the Upper Cretaceous of central east Sinai was controlled by a long-term transgressive phase and several higher order sea level fluctuations. In the study, six different facies zones have been differentiated. The main depositional elements include coastal mudflats with tidal channels, gypsiferous sabkha plains, peritidal siliciclastics, peritidal carbonates, high and low energy carbonate inner shelf facies as well as microfossil-rich outer shelf pelites. The paper gives a first sequence stratigraphic interpretation for this interval in the region based on detailed sedimentologic, biostratigraphic and paleoecologic investigations in 13 Turonian-Maastrichtian sections and a review of all available data from the literature. Biostratigraphy is mainly based on planktonic foraminifera, calcareous nannofossils, ostracods and ammonites. The study is restricted to an area which was tectonically rather quiet during the Late Cretaceous-Early Tertiary lying south of the Syrian Arc intraplate foldbelt which experienced major uplifting during this period. Within the the Turonian to Maastrichtian interval, six major sequence boundaries have been reconstructed. Cycle duration varies between 4 to 9 Ma which points to a cycle order being intermediate between 3rd and 2nd. Correlation with other sea level reconstructions from the region (Egypt, Israel, Jordan, Tunisia) points to a more or less synchronous regional sea level development. Comparison of the regional sequences with the 'eustatic' model of Haq et al. (1987) involves uncertainties; nevertheless some of the sea level fluctuations recorded in Sinai may be attributed to worldwide eustatic sea level changes.
6. Late Maastrichtian litho- and ecocycles from the hemipelagic of Eastern Sinai, Egypt

S. Lüning, A. M. Marzouk, J. Kuss

a University of Bremen, FB5 - Geosciences, P. O. Box 330440, 28334 Bremen, Germany.
b Tanta University, Faculty of Science, Geology Dept., Tanta 31511, Egypt

Abstract

A detailed biostratigraphic and paleologic study by means of calcareous nannofossils and planktonic and benthonic foraminifera has been carried out in four sections of hemipelagic marls and chalks of the Late Maastrichtian Abathomphalus mayaroensis Zone of Eastern Sinai in order to evaluate the mechanisms controlling the composition of the well preserved microfauna and flora. The Abathomphalus mayaroensis Zone in Eastern Sinai can be easily identified by the wide occurrence of the index fossil A. mayaroensis and can be further subdivided by the first occurrences of Plummerita reicheli (ex. P. hantkeninoides) and Mkula prinsii. A higher resolving cyclostratigraphic subdivision of the interval based on a low-frequency cyclicity in the foraminiferal planktonic benthonic (P/B) ratio is proposed. Microfossil abundances and lithologies are characterized by pronounced repetitive distribution patterns. This includes low- and high-frequency P/B fluctuations, repetitive changes in the abundance of calcareous nannofossils and benthonic foraminifera, as well as the development of chalk marl couplets and thinning-upward chalk bundles. Both microfossil distribution patterns and rhythms are attributed to changes in paleoproductivity. Semiquantitative investigations of calcareous nannofossils and benthonic foraminifera allowed to differentiate a high (HP) and low (LP) productivity assemblage. While the HP assemblage is dominated by Glaukolithus diplogo grammus, Manvitella pemmatoida, Microrhabdulus decoratus and Micula muns and the benthonic foraminifera Neoflabellina jarvisi, the LP assemblage is characterized by Lithraphidites quadratus and Bolivinoides draco. The chalk marl couplets, thinning-upward chalk bundles and the high-frequency P/B patterns are interpreted to reflect productivity changes related to orbital forcing. The studied hemipelagites have obviously been deposited during the latest phase of the southern Tethyan upwelling system, which was active from the Santonian to the Late Maastrichtian with a peak in the Campanian (Almogi-Labin et al., 1993). Termination of upwelling just before the K/T boundary also provides a good explanation for the change towards a paleobathymetric control on foraminiferal distribution as observed for the Paleocene of Central East Sinai (cf. Lüning et al., submitted b).

6.1. Introduction

The analysis of periodic and episodic repetitive processes forms the basis for the reconstruction of many geologic processes. Cyclicalities occur on every scale, and can be measured by means of numerous parameters (e.g. Weller, 1964; Duff et al., 1967; Einsele & Seilacher, 1982; de Boer & Smith, 1994). This contribution is focused on variations in calcareous nanofossil and benthonic foraminifunal assemblages in a hemipelagic setting during the Late Maastrichtian Abathomphalus mayaroensis Zone. Control mechanisms are complex, nevertheless sea level and paleoclimate are the most likely agents in shaping the foraminiferal fauna (Bé, 1977; Boltovskoy & Wright, 1976; Murray, 1973, 1991; Vincent & Berger, 1981; Gibson, 1989, and Van der Zwaan et al., 1990). If the foraminiferal assemblages are predominantly controlled by paleobathymetry, suitable indices like the foraminiferal planktonic benthonic ratio (P/B) can be used as a proxy for sea-level changes, which may contribute significantly to sequence stratigraphic reconstructions. While previous investigations of the Paleocene in the study area in East Sinai have shown that the P/B curve can be directly related to sea-level changes in most cases (Lüning et al., submitted b), this relationship had to be tested for the Late Maastrichtian.

First evidence for an involvement of climatically induced paleoproductivity changes comes from the development of chalk-marl couples within the studied series. Similar hemipelagic / pelagic limestone-marl alternations from the middle Cretaceous to Paleocene are described from many areas (Leary et al., 1989; Herbert & D'Hondt, 1990; Huang et al., 1992; Ten Kate & Sprenger, 1992; Park et al., 1993; Noé, 1993; Bellanca et al., 1996). In most cases, the couples are attributed to the orbital precessional signal with a mean period of 20 ky and the chalks or limestones are interpreted to be associated with paleoclimatically induced high productivity (Herbert et al., 1995; R.O.C.C., 1986). The rhythms described here for East Sinai have already been geochemically studied in West Sinai by Abdel Fattah et al. (1996a,b) who also attribute the chalk-marl succession to orbital forcing.
Due to this evidence for a paleoclimatic control on Late Maastrichtian lithofacies, a closer examination of the highly variable nannofossil and foraminiferal assemblages seemed necessary in order to learn more about their control mechanisms and to avoid potential paleobathymetric misinterpretations. Good descriptions for an orbitally induced paleoclimatic control on Cretaceous foraminifera, for example, have already been presented from Southern England (Leary et al., 1989; Leary & Hart, 1992) and from Central Italy (Premoli Silva et al., 1989).

For Northeast Africa / Northern Arabia, detailed investigations on foraminifera, calcareous nannofossils and dinoflagellate cysts of two Campanian to Maastrichtian subsurface sections in Israel (Almogi-Labin et al., 1990b; Almogi-Labin et al., 1993; Eshet et al., 1994; Eshet & Almogi-Labin, 1996) suggest that paleoproductivity is a dominant factor in shaping the microfossil assemblages during this period.

6.2. Regional Setting

During Maastrichtian times, Sinai was located on the broad northern shelf margin of the Afroarabian Plate. The Gulf of Suez and Gulf of Aqaba rifts, which bound the Sinai microplate today, were still closed. While Northern Sinai and the adjacent Negev were affected by transpressive movements from the Turonian onwards, Central and Southern Sinai are believed to have remained tectonically rather quiet throughout the whole Mesozoic and Early Tertiary (Said, 1962; Cohen et al., 1990; Kerdany & Cherif, 1990). The SE-vergent, NE-SW trending domal anticlines in Northern Sinai are part of the 'Syrian Arc' (Krenkel, 1924, 1925) which represents an intraplate foldbelt extending from Egypt to Syria and which was formed by Late Cretaceous to recent inversion of Late Triassic / Liassic halfgrabens (Moustafa & Khalil, 1990, Chaimov et al., 1993, Shahar, 1994). Following the nomenclature of Said (1962), the inversion zone is also termed 'unstable shelf', and the southern tectonically calm block 'stable shelf' (Fig. 6-1). The sections studied here are situated on both the stable and unstable shelves.

During the studied Maastrichtian interval, Central Sinai was part of a northeast-southwest striking intrashelf basin that was confined on the S by the Afroarabian paleoshoreline in the area of the modern Red Sea and on the N by the domal anticlinal / synclinal NE-SW striking belts belonging to the Syrian Foldbelt (paleogeographic, paleobathymetric and isopach maps in Bartov & Steinitz, 1977; Sestini, 1984; Camoin et al., 1993; Kuss & Bachmann, 1996). Studies of foraminifera showed that during the Maastrichtian, a middle neritic to upper bathyal environment prevailed in large parts of Central Sinai (Abbas et al., 1994; Shahin, 1990) and Northern Sinai / Negev (Hewaidy & El Ashwah, 1993; Almogi-Labin et al., 1990a). Shallower facies and locally subaerial exposure are assumed for some of the anticlinal highs in Northern Sinai / Negev Desert (Almogi-Labin et al., 1990a). The Late Maastrichtian hemipelagites are composed of marly foraminiferal chalks which are classified as 'Sudr Chalk' or 'Sudr Formation' in Sinai (Ghorab, 1961; Ziko et al., 1993) and as 'Ghrehab Formation' in the Negev (Bartov et al., 1972). According to Reiss (1988), Shemesh & Kolodny (1988) and Almogi-Labin et al. (1993) these sediments were accumulated in a highly productive upwelling regime that dominated the southern Tethys margin from the Santonian to Late Maastrichtian.

Until now, there have been only few investigations of foraminifera and calcareous nannofossils dealing with the Maastrichtian hemipelagites of Sinai. While the Late Cretaceous to Early Eocene biostratigraphic framework of Sinai has been described e.g. by Nakkady (1957), Abdelmalik et al. (1978), Hewaidy et al. (1991) and Marzouk & Hussein (1994), paleocologic studies including the Maastrichtian of Sinai have been presented by Hewaidy & El Ashwah (1993), Ismail (1992), Abbas et al. (1994), Anan (1992) & Shahin (1992).

6.3. Materials and Methods

During two field expeditions in 1995/96, a total of 27 sections from the Upper Cretaceous-Lower Tertiary of Central East and North East Sinai were newly measured, of which six sections from the Late Maastrichtian are
used for this contribution (Figs. 6-1, 6-2). All six sections were sampled at intervals ranging between 1-3 metres. To isolate the foraminifera from the matrix, approximately 500g of each of the 84 samples were washed twice with a 63μm sieve after treatment with H₂O₂ and the highly concentrated tenside REWOQUAT, respectively. The foraminiferal planktonic benthonic (P/B) ratio was routinely determined for all samples by counting traverses with a minimum of 300 specimens in the unfractonated washing residue. The value is calculated in percent plankton or benthos in relation to the total number of counted foraminifera. The microfossil residue was then fractionated into four grain size classes for easier handling. The planktonic foraminifera were identified under the light microscope (by Lüning) using the atlases and zonal concepts of Caron (1985) and Robaszynski et al. (1984).

Determination and paleobathymetrical interpretation of the benthonic foraminifera were done (by Lüning) after illustrations and concepts of Aubert & Berggren (1976), Berggren & Aubert (1975), Luger (1985), LeRoy (1953), Said & Kenawy (1956), Saint-Marc (1992), Speijer (1994a and pers. comm.) and Van Morkhoven et al. (1986). Detailed systematic descriptions can be found in these papers as well and are therefore not repeated here. For semiquantitative investigations, the whole washing residue of the fraction >250 μm was studied in 66 samples. Some of the studied benthonic species also occurred in the fraction >125 μm, but since their quantities were small and their abundance was directly related to that of the coarser fraction, the smaller fraction was ignored. From the fraction >250 μm, all representatives of the selected species were picked and counted. A similar method has previously been successfully employed by Leary et al. (1989). Data was converted into abundance classes (none / 0 specimen, rare / 1, few / 2-4, common / 5-10, abundant / >10, very abundant / >100).

For the calcareous nannoplankton, smear slides were prepared using techniques described in Bramlette & Sullivan (1961) and Hay (1961, 1965). The slides were examined (by Marzouk) under the light microscope at a magnification of about x1250 by both cross-polarized and phase-contrast. Biostratigraphy is based on the zonal concept of Thierstein (1976) and Monechi & Thierstein (1985). Relative abundance classes of the calcareous nannofossils are based on the system: rare = 1 specimen in 200 fields of view, few = 1/50, common = 1/10, and abundant = 1/1.

6.4. Biostratigraphy

The present study concentrates on the Late Maastrichtian Abathomphalus mayaroensis Zone (Fig. 6-2) which is defined as the interval from the first occurrence (FO) of the planktonic foraminifera A. mayaroensis (Bolli) to the first occurrence of Paleocene foraminifera (Brönnimann, 1952; Caron, 1985). In Sinai, the zonal marker species is widely distributed, so that the biozone can be easily identified. This contrasts the situation in other areas such as Southern Egypt, where A. mayaroensis is missing in age-equivalent sediments (e.g. Luger, 1985). The A. mayaroensis biozone can be further subdivided by the FO of Plummerita reicheli (Brönnimann) (sensu Masters, 1993), a characteristically small planktonic form with radially elongate, turbulospinate chambers. This species is also present in areas where A. mayaroensis is absent, e.g. in Southern Egypt (Luger et al., 1997, in press; discussion of biozonation herein). Another biostratigraphic datum within the A. mayaroensis Zone is provided by the FO of the calcareous nannofossil species Micula prinsii. In the sections from Northeast Sinai (A1, A5, A7) the FO of P. reicheli was found to be slightly before the FO of M. prinsii (Fig. 6-2). However, in Central Sinai, P. reicheli seemed to occur less frequently and a virtually inverse relationship was recorded (sections F, P). The latter may be due to paleoecologic reasons so that the FO datum of P. reicheli should be used with caution. Since P. reicheli was often found together with A. mayaroensis, the idea of Masters (1993) that P. reicheli occurs after the extinction of A. mayaroensis, cannot be supported by our material.

6.5. Paleoeologic Parameters Studied

Planktonic Benthonic Foraminiferal Ratio and Benthonic Foraminifera

Productivity of planktonic and benthonic foraminifera is influenced by numerous water mass properties such as temperature, pressure, density, nutrients, salinity, light penetration, oxygen, as well as other physical, chemical and biological factors. The two main important processes capable of affecting these parameters are sea-level changes and climatic / oceanographic fluctuations. If the water mass properties are mainly controlled by water
depth, physiobiocchemical parameters and therefore the distribution patterns of planktonic and benthonic foraminifera often vary predictably along a depth gradient. One of the major changes represents the increase in the percentage of planktonic foraminifera in the bottom assemblages at greater depths. This reflects the trend that planktonic species occur in greater densities in open oceanic environments, whereas the abundance of benthonic foraminifera is higher in neritic environments (Phleger, 1964; Reiss et al., 1974; Gibson, 1989; Van der Zwaan et al., 1990). An increase in the foraminiferal planktonic benthonic (P/B) ratio may therefore be interpreted in a suitable environment as a period of deepening. Abundance of certain benthonic foraminiferal species can also be linked to paleodepth and may therefore be used for approximate paleobathymetric trend analyses (Murray, 1991; Speijer 1994a,b). Previous studies have shown that many species and genera of benthonic foraminifera can be attributed to distinct broad depth ranges (Berggren & Aubert, 1975; Aubert & Berggren, 1976; Van Morkhoven et al., 1986; Saint-Marc, 1992; Speijer, 1994a; Luger, 1985) (Tab. 6-1).

If the water mass properties are dominated by climatically induced productivity changes (e.g. upwelling or sinking of saline, oxygen-poor, nutrient-rich water plumes), different trends in foraminiferal distribution are developed (e.g. Altenbach & Sarntheim, 1989; Widmark, 1995). In this case, the P/B ratio may be sensitive for productivity changes. A good example is provided by Leary et al. (1989) and Leary & Hart (1992) who recorded P/B changes of up to 50% across a paleoclimatically controlled chalk-marl couplet of the Cenomanian of Southern England. They demonstrated that the highest planktonic values were reached in chalks which they interpreted as having been deposited during climatically induced surface-water productivity blooms. Productivity fluctuations in the surface waters may also lead to significant changes in the benthonic foraminifera fauna. Leary et al. (1989) for example postulate that substrate changes were caused by climatically-induced surface water productivity blooms, which in turn may lead to a reorganisation of the benthonic foraminiferal assemblages. High productivity events may also cause oxygen depletion with a bloom of anaerobic forms (Kaiho, 1991).


<table>
<thead>
<tr>
<th>Bathymetric zonation after a combination of concepts used by Bassioni &amp; Lüger (1980) and Sitter &amp; Baker (1972)</th>
<th>SHELF</th>
<th>SLOPE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bathymetric zonation of Paleocene benthonic foraminifera after Berggren &amp; Aubert (1975)</td>
<td>Tethys Carbonate Fauna (TCF)</td>
<td>Midway Type (MF)</td>
</tr>
<tr>
<td>Bathymetric zonation after a combination of concepts used by Van Morkhoven et al. (1986) and Saint-Marc (1992)</td>
<td>neritic</td>
<td>bathyal</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>species</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bolivinoides draco (Marsson)</td>
<td>Van Morkhoven et al. (1986)</td>
</tr>
<tr>
<td>Dorothea oxycona (Reuss)</td>
<td>Saint-Marc (1992)</td>
</tr>
<tr>
<td>Frondicularia spp.</td>
<td>Speijer pers. comm. (1966)</td>
</tr>
<tr>
<td>Neoflabellina jarvisi (Cushman)</td>
<td>Olsson &amp; Nyong (1964)</td>
</tr>
<tr>
<td>Neoflabellina semireticulata (Cushman &amp; Jarvis)</td>
<td>Berggren &amp; Aubert (1975)</td>
</tr>
<tr>
<td>Stensioina excolata (Cushman)</td>
<td>Saint-Marc (1992)</td>
</tr>
<tr>
<td>Tritaxia midwayensis (Cushman)</td>
<td>Van Morkhoven et al. (1986)</td>
</tr>
</tbody>
</table>

Table 6-1. Typical paleobathymetric ranges for the semiquantitatively studied species after literature data. Note especially the bathyal / abyssal character of D. oxycona and N. semireticulata in comparison to the inner neritic preference of S. knebeli.
**Calcareous Nannofossils**

Calcareous nannofossils are mainly used in biostratigraphy rather than as paleoenvironmental indicators due to their planktic mode of life and rapid distribution. Under certain conditions their distribution patterns may reflect the ecological conditions within the surface waters (Murray, 1995, Eshet et al., 1994; Eshet & Almogi-Labin, 1996). Nevertheless, sea-level reconstructions by means of calcareous nannofossils (e.g. Reale, 1995) are complicated (Perch-Nielsen, 1985).

**6.6. Results**

The study of the Late Maastrichtian hemipelagites of Eastern Sinai showed that microfossil contents and lithologies are characterized by pronounced repetitive distribution patterns. This includes low- and high-frequency P/B fluctuations, repetitive changes in the abundance of calcareous nannofossils and benthonic foraminifera, as well as the development of chalk marl couplets and thinning-upward chalk bundles.

**P/B-Cyclicity**

Within the studied sections, a low- and a high-frequency cyclicity in the P/B ratio was found. While the long-periodic P/B curves (sections A7 and P) are quite similar in appearance, the short-periodic P/B curves differ markedly, with a high amplitude signal in section F and an only small amplitude in section A5 (Fig. 6-2). It has to be noted that frequency distribution in the different sections is rather independent of the geographic grouping (Fig. 6-1). The general trend of the short-periodic P/B pattern in section F suggests that the high frequency signals are superimposed on the longperiodic P/B pattern. Based on this observation, an idealized regional P/B cycle has been compiled. A number of characteristic maxima (H, high) and minima (L, low) in the curves were selected for a cyclostratigraphic correlation (Fig. 6-2). The idealized P/B cycle is best represented in section A7. Since the studied sections are located on both the unstable and the stable shelf of Northern and Central Eastern Sinai, the cycles can be cross-correlated independently of the tectonic setting across Sinai and may represent a valuable cyclo- and ecostratigraphic tool (e.g. Perlmutter & Matthews, 1989). Comparable age-equivalent fluctuations in the planktonic-benthonic distribution can be also found in descriptions of sections from Southwest Sinai (Shahin, 1990) and Northeast Sinai (Hewaidy & El Ashwah, 1993).

**Benthonic Foraminifera**

The interval studied here yields a rich, highly diverse, well preserved benthonic foraminiferal fauna (Tab. 6-2). Major solution effects or matrix fragments after washing were not found except in one case (marked by 'pr' in section A7). Taphonomic effects in the distribution patterns seem therefore unlikely. A comprehensive systematic description of this highly diverse outer neritic-upper bathyal Tethyan fauna was already published in the first half of this century in various detailed papers, mainly by Cushman and co-workers (e.g. Cushman, 1946; Cushman & Renz, 1946). For paleologic reconstructions, a semiquantitative study of seven selected benthonic foraminiferal species [Bolivinoides draco (Marsson), Dorothia oxyconia (Reuss), Neoflabellina jarvisi (Cushman), Neoflabellina semireticulata (Cushman & Jarvis), Spiroplectammina knebeli Le Roy (~Sp. laevis Cushman), Stensioina excolata (Cushman), rounded form of Trithixia midwayensis (Cushman)] and one undifferentiated genus (Frondicularia spp.) was carried out for 66 samples from four sections (A5, A7, F, P) (Plate 6-1). Criteria for the selection of the species were a) availability of paleobathymetric data from the literature (Tab. 6-1) and/or b) characteristic, easily determinable form.

**Figures 6-3, 6-4. Distribution charts for calcareous nannofossils and benthonic foraminifera with P/B curves at left (compare to Fig. 6-2). HighProd / LowProd indicate intervals of high and low surface water productivities, as interpreted from the microfossil distribution patterns. HP / LP mark the high and low productivity assemblages. Shaded bars emphasize abundance trends for selected marker species, which positively match the model in Fig. 6-8 with regard to the assumed high or low productivity phases. Preservation ranges from good (solid boxes) to moderate (ovals).**

**Figure 6-5. Semiquantitative abundances of four selected benthonic foraminiferal species (bars and lines) in relation to the respective P/B ratios (curves and dots) within the cyclostratigraphic frame (idealized P/B curve at left derived from Fig. 6-2). The abundance charts exhibit a complex and often cyclical distribution for most of the studied species. However, a regionally synchronous abundance development can be partly interpreted, which then is plotted into the idealized cyclostratigraphic P/B curve in the leftmost box for each species. Legend: pr = sample with poor preservation, some of the abundances therefore questionable; bis = planktonic foraminifera consist predominantly of biserial forms, most probably associated with episodes of high productivity.**
Tritaxia midwayensis (rounded form considered only)

Neoflabellina semireticulata

Frondicularia spp.
group. The second assemblage (marked as LP in Figs. 6-3, 6-4) consists of the Lithraphidites quadratus and the benthonic foraminifera Bolivinoides draco. Highest abundances of these species are in general associated with high planktonic values. A similar relationship can be observed for the pair St. excolata, Neoflabellina semireticulata, Osangularia plummerae Broten, and Nodosaria longiscala (D’Orbigny).

The study showed that within the studied interval, abundances of the investigated benthonic species change significantly in a repetitive manner (Figs. 6-3, 6-4, 6-5). Various frequency / amplitude patterns are observed, including for example high frequency / strong amplitude (e.g. St. excolata in sect. F, Figs. 6-4, 6-5), low frequency / strong amplitude (Tr. midwayensis in sect. A5, Figs. 6-3, 6-5) and low frequency / weak amplitude (N. semireticulata in sect. A7, Figs. 6-3, 6-5). Abundance transitions vary from gradual (e.g. Tr. midwayensis in sect. F, Figs. 6-4, 6-5) to abrupt (e.g. Bo. draco in sect. P, Fig. 6-4). One species (Sp. knebeli) is often completely missing or is rare with only a few but significant exceptions (Figs. 6-3, 6-4). Within the cyclostratigraphic frame (see above and Fig. 6-2), a comparison of age-equivalent abundance patterns for individual species in the four studied sections was attempted. It seems that some of the abundance fluctuations can be correlated peak-by-peak with patterns developed in the neighbouring sections (Fig. 6-5). For these cases, an idealized abundance curve was reconstructed. However, correlatability varies significantly from species to species. Whereas for some species, regional abundance patterns can be more easily reconstructed (e.g. T. midwayensis, St. excolata), for other groups local, non-basin wide abundance trends seem to dominate (e.g. Sp. knebeli).

A comparison between the abundances of different species yields only a few trends which can be generalized. Among these are the distribution patterns of D. oxycona and Frondicularia spp. which show a high degree of correspondence in all sections (Figs. 6-3, 6-4). Parallel distribution patterns have also been observed for the pair St. excolata and D. oxycona in section F (Fig. 6-4) and for St. excolata and N. semireticulata in section A7 (Fig. 6-3). An inverse development in abundance can often be found for St. excolata and Sp. knebeli in all sections studied (Figs. 6-3, 6-4).

The comparison of the abundances of benthonic foraminiferal species with the P/B ratios yields also only few results which can be generalized. Highest abundances of N. semireticulata, for example, are often associated with high planktonic values. A similar relationship can be observed for St. excolata around the cyclostratigraphic level 2L (Fig. 6-5). Within the high-frequency P/B cycle of section F, highest abundances of D. oxycona are reached during periods of high planktonic values (Fig. 6-4).

Calcareous Nannofossils (Plate 6-2)

Abundance distribution of most studied calcareous nannofossil species was also found to be highly fluctuating (Figs. 6-3, 6-4). A detailed comparison of the individual distribution patterns of the calcareous nannofossils and the selected benthonic foraminifera suggests that two groups of species characterized by an often inverse abundance development can be distinguished. The first assemblage (marked as HP in Figs. 6-3, 6-4) consists of Glaukolithus diplogrammus, Manvitella pemmatoides, Microrhahdulus decoratus and Micula murus and the benthonic foraminifera Neoflabellina jarvisi. Within this group, the individual distribution patterns in general show a high degree of correspondence. Furthermore, it has been found that rare blooms of biserial planktonic foraminifera, including heterohelicids or buliminids are associated with high abundances in the first assemblage group. The second assemblage (marked as LP in Figs. 6-3, 6-4) consists of Lithraphidites quadratus and the benthonic foraminifera Bolivinoides draco. Highest abundances of these species are in general associated with low abundances of representatives of the first assemblage group.

---

Table 6-2. List of identified benthonic foraminifera from the lower part of the Gebel Misheiti section (F).

<table>
<thead>
<tr>
<th>Foraminifera</th>
<th>Identification Details</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alabamina obtusa (Burrows &amp; Holland)</td>
<td></td>
</tr>
<tr>
<td>Alabamina wilcoensis Toulmin</td>
<td></td>
</tr>
<tr>
<td>Ammodiscus glabratus Cushman &amp; Jarvis</td>
<td></td>
</tr>
<tr>
<td>Anomalinaeoides acuta (Plummer)</td>
<td></td>
</tr>
<tr>
<td>Anomalinaeoides michaelwensis (Plummer)</td>
<td></td>
</tr>
<tr>
<td>Astococus bifurcatus LeRoy</td>
<td></td>
</tr>
<tr>
<td>Bolivinoides draco (Marsson)</td>
<td></td>
</tr>
<tr>
<td>Cibicides beaudelli LeRoy</td>
<td></td>
</tr>
<tr>
<td>Cibicides aleeni (Plummer)</td>
<td></td>
</tr>
<tr>
<td>Conophastoma incrassata (Reuss)</td>
<td></td>
</tr>
<tr>
<td>Dentalina coeli Cushman &amp; Dusenbury</td>
<td></td>
</tr>
<tr>
<td>Dorothia oxycona (Reuss)</td>
<td></td>
</tr>
<tr>
<td>Eponoides plummerae Cushman</td>
<td></td>
</tr>
<tr>
<td>Frondicularia cf. frankel Cushman</td>
<td></td>
</tr>
<tr>
<td>Frondicularia spp.</td>
<td></td>
</tr>
<tr>
<td>Gyroidinoides girardamus (Reuss)</td>
<td></td>
</tr>
<tr>
<td>Lagena hispida Reuss</td>
<td></td>
</tr>
<tr>
<td>Lenticulina oligostega (Reuss)</td>
<td></td>
</tr>
<tr>
<td>Lenticulina sp.</td>
<td></td>
</tr>
<tr>
<td>Neoflabellina jarvisi (Cushman)</td>
<td></td>
</tr>
<tr>
<td>Neoflabellina semireticulata (Cushman &amp; Jarvis)</td>
<td></td>
</tr>
<tr>
<td>Nodosaria latejugata group, Gumbel</td>
<td></td>
</tr>
<tr>
<td>Nodosaria longiscala (D’Orbigny)</td>
<td></td>
</tr>
<tr>
<td>Nuttallides triumpyi (Nuttall)</td>
<td></td>
</tr>
<tr>
<td>Osangularia plummerae Broten</td>
<td></td>
</tr>
<tr>
<td>Robulus tisdis (Schwager)</td>
<td></td>
</tr>
<tr>
<td>Saracenaria saratogana Howe &amp; Wallace</td>
<td></td>
</tr>
<tr>
<td>Sitella colonensis (Cushman &amp; Hedberg)</td>
<td></td>
</tr>
<tr>
<td>Spiroplectammina knebeli LeRoy</td>
<td></td>
</tr>
<tr>
<td>Stensioina excolata (Cushman)</td>
<td></td>
</tr>
<tr>
<td>Stiltomella paleocenica (Cushman &amp; Todd)</td>
<td></td>
</tr>
<tr>
<td>Trisactia midwayensis (Cushman)</td>
<td></td>
</tr>
<tr>
<td>Vagimidopsis longiforma (Plummer)</td>
<td></td>
</tr>
<tr>
<td>Vaginulina sp.</td>
<td></td>
</tr>
<tr>
<td>Valvulineria aegyptiaca LeRoy</td>
<td></td>
</tr>
</tbody>
</table>

---

The study showed that within the studied interval, abundances of the investigated benthonic species change significantly in a repetitive manner (Figs. 6-3, 6-4, 6-5). Various frequency / amplitude patterns are observed, including for example high frequency / strong amplitude (e.g. St. excolata in sect. F, Figs. 6-4, 6-5), low frequency / strong amplitude (Tr. midwayensis in sect. A5, Figs. 6-3, 6-5) and low frequency / weak amplitude (N. semireticulata in sect. A7, Figs. 6-3, 6-5). Abundance transitions vary from gradual (e.g. Tr. midwayensis in sect. F, Figs. 6-4, 6-5) to abrupt (e.g. Bo. draco in sect. P, Fig. 6-4). One species (Sp. knebeli) is often completely missing or is rare with only a few but significant exceptions (Figs. 6-3, 6-4). Within the cyclostratigraphic frame (see above and Fig. 6-2), a comparison of age-equivalent abundance patterns for individual species in the four studied sections was attempted. It seems that some of the abundance fluctuations can be correlated peak-by-peak with patterns developed in the neighbouring sections (Fig. 6-5). For these cases, an idealized abundance curve was reconstructed. However, correlatability varies significantly from species to species. Whereas for some species, regional abundance patterns can be more easily reconstructed (e.g. T. midwayensis, St. excolata), for other groups local, non-basin wide abundance trends seem to dominate (e.g. Sp. knebeli).

A comparison between the abundances of different species yields only a few trends which can be generalized. Among these are the distribution patterns of D. oxycona and Frondicularia spp. which show a high degree of correspondence in all sections (Figs. 6-3, 6-4). Parallel distribution patterns have also been observed for the pair St. excolata and D. oxycona in section F (Fig. 6-4) and for St. excolata and N. semireticulata in section A7 (Fig. 6-3). An inverse development in abundance can often be found for St. excolata and Sp. knebeli in all sections studied (Figs. 6-3, 6-4).

The comparison of the abundances of benthonic foraminiferal species with the P/B ratios yields also only few results which can be generalized. Highest abundances of N. semireticulata, for example, are often associated with high planktonic values. A similar relationship can be observed for St. excolata around the cyclostratigraphic level 2L (Fig. 6-5). Within the high-frequency P/B cycle of section F, highest abundances of D. oxycona are reached during periods of high planktonic values (Fig. 6-4).
PLATE 6-1. Semiquantitatively studied benthonic foraminifera and biostratigraphically important planktonic foraminifera. All scale bars = 100 μm (except No. 6-9: scale bar = 300 μm).
2. Dorothia oxyconia (Reuss), section A5, sample A5-14.
4. Tritaxia midwayensis (Cushman) (rounded form), section A5, sample A5-3.
5. Frounticulata spp., section P, specimens 6 and 7 from sample P1-11, specimen 8 from sample P1-7, specimen 9 from sample P1-6.
6-9. Stensioeina excolata (Cushman), section A5, all specimens from sample A5-3.
10-12. Nuttallides truempyi (Nuttall), section P, specimen 13 from sample P1-11, specimen 14 from sample P1-10.
15-16. Plummerita reicheli (Brönnimann) [sensu Masters, 1993], section A5, both specimens from sample A5-17.
19. Neoflabellina jarvii (Cushman), section A5, both specimens from sample A5-4.

PLATE 6-2. Calcareous nannofossils: species of the high and low productivity assemblages and other important forms. Scale bar = 15 μm.
3-4. Arkhangelskiella cymbiformis Vekshina, section A5, sample A5-17.
11. Micula staurophora (Gardet) Stradner, section F, sample F1-2.
A series of well-developed chalk-marl couplets on a decimetre scale with a total thickness of at least 21 m was found in the uppermost Maastrichtian at Gebel Misheiti (section F, East Central Sinai, 10 km NNE' Themed) (Figs. 6-6, 6-7). Similar bedding rhythms are developed with varying intensities throughout the study area. The marls at Gebel Misheiti are strongly bioturbated and consist of a mixture of fine-grained, siliciclastic clay/silt particles and calcareous hemipelagic microorganisms, mainly calcareous nannofossils and foraminifera. The beds are of remarkable constant thickness (7-9 cm). The relatively hard, almost entirely biogenic chalk beds, in contrast, are arranged in thinning-upward bundles. It is clear that the number of chalk beds exposed at Gebel Misheiti is too small to allow statistically backed investigations. However, it seems that an individual thinning-upward bundle consists of 4 to 6 chalk beds (Fig. 6-7). It can be observed that the average thickness of a bundle matches the thickness of a single high-frequency P/B cycle. The two uppermost cycles in the P/B-curve in Fig. 6-7 seem to correspond to the two thinning-upward cycles recorded in the same interval.

Figure 6-6 (at right). Precessional chalk-marl rhythms at Gebel Misheiti in the Late Maastrichtian A. mavaroensis Zone of Central East Sinai (see also Fig. 6-7). Chalks appear in light colours, marls are darker. Backpack at base for scale. Note the thickness changes of the chalk beds which contrast the constant thicknesses of the marl interlayers.
thinning-upward chalk bundles suggests that the marl layers have been formed during times with a rather constant cycle. Furthermore, the typical stacking pattern at Gebel Misheiti with marl layers of constant thickness and upward chalk bundles of 4-6 beds are consequently interpreted as being controlled by the 100 ky eccentricity to apply the same interpretation of orbital precessional control to the chalk-marl couplets in Sinai. The thinning-precessional cycle (Herbert et al., 1995). Due to lithological analogy and stacking reasons, it seems reasonable sites of Late Cretaceous to Early Paleocene age display similar cycles whose mean period, estimated by paleomagnetically determined sedimentation rates, is close to 20 ky, the expected mean period of the settings (R.O.C.C, 1986; Botter & Almogi-Labin, 1993). A large number of Deep Sea Drilling Project drill 'hulk-marl couplets (European Late Cretaceous carbonate sequences deposited in hemipelagic epicontinental and continental-edge settings). Rhythmic bedding on a decimetre scale is a prominent feature frequently encountered in North American and European Late Cretaceous carbonate sequences deposited in hemipelagic epicontinental and continental-edge settings (R.O.C.C., 1986; Botter et al., 1986; Fischer, 1993). A number of Deep Sea Drilling Project drill sites of Late Cretaceous to Early Paleocene age display similar cycles whose mean period, estimated by paleomagnetically determined sedimentation rates, is close to 20 ky, the expected mean period of the precessional cycle (Herbert et al., 1995). Due to lithological analogy and stacking reasons, it seems reasonable to apply the same interpretation of orbital precessional control to the chalk-marl couplets in Sinai. The thinning-upward chalk bundles of 4-6 beds are consequently interpreted as being controlled by the 100 ky eccentricity cycle. Furthermore, the typical stacking pattern at Gebel Misheiti with marl layers of constant thickness and thinning-upward chalk bundles suggests that the marl layers have been formed during times with a rather constant
sedimentation rate, and that the chalks have been deposited during periods of elevated calcareous plankton production, diluting the background clay sedimentation. Biogenic production events may have been controlled by the orbital 20 ky precessional signal, leading to the periodic, equidistant intercalation of chalks into the marls. Similar orbitally plankton productivity models for chalks and limestones in comparable Cretaceous rhythmic successions have been presented by several authors (e.g. de Boer, 1982; Herbert & Fischer, 1986; Ditchfield & Marshall, 1989; Bellanca et al., 1996). The stacking pattern described here for Sinai provides a good field example for a theoretical model by Lüning & Ricken (1991: Fig 7a) for limestone-marl alternations controlled by carbonate productivity, although in contrast, they assume a thickening upwards trend.

Abdel Fattah et al. (1996a+b) investigated a similar Maastrichtian rhythmic succession in West Central Sinai and also interpreted the couplets as being associated with the precessional Milankovitch signal. Their model is based on detailed geochemical and mineralogical studies of 103 couplets. According to these authors, the chalk beds represent biogenic productivity intervals during drier climates, which corresponds to our interpretation. In contrast to us, they suggest that the marls are dilution products deposited during humid times with increased terrestrial runoff and exaggerated influx of detrital clays. This seems unlikely here because of the rather constant thickness of the marl layers found in Central East Sinai.

The fact that an average thinning-upward cycle and a single high-frequency P/B cycle possess similar thicknesses (Fig. 6-7), suggests that the high-frequency P/B cyclicity may be related to the 100 ky eccentricity signal as well. A similar link between lithological cycles and foraminifer abundance cycles has previously been presented from the Cenomanian of Southern England (Leary et al., 1989; Leary & Hart, 1992; Paul, 1992) who recorded P/B changes of up to 50% across a precessationally controlled chalk-marl couplet. They were able to show that the highest planktonic values were reached in the chalks which they interpreted as having been deposited during climatically induced surface-water productivity blooms. A similar productivity mechanism is supposed here for the Late Maastrichtian in Eastern Sinai, as based on indirect evidence from the correlation of chalk bundles with the high-frequency P/B cyclicity and from the association of the low productivity assemblage with benthonic abundance peaks of the low-frequency P/B pattern.

Whether the low-frequency P/B patterns reflect the 400 ky eccentricity signal or are random, must remain speculation. The two P/B cycles within the A. mayaroensis Zone, which has a duration of approx. 2 my (Gradstein et al., 1995, Braquier et al., 1995) seem to be significantly shorter than 1 my, due to a hiatus, an interval of reduced sedimentation and probably loss of cyclicity in the upper part of the zone (cf. Lüning et al., submitted b). That eccentricity may affect the distribution of microfossils was shown in quantitative investigations by Haas et al. (1994), who studied radiolarians and calpionellids from the Lower Cretaceous of Hungary. They demonstrated that the temporal microfossil distribution patterns were mainly controlled by the 400 ky eccentricity signal. Beaudoin et al. (1995) studied a Cenomanian-Turonian boundary section in Colorado, and also attribute the observed microfaunal changes to eccentricity.

What do foraminifera contribute to sea-level reconstructions in the Late Maastrichtian of Eastern Sinai?

In order to determine whether the effects of sea-level changes can be detected within the foraminifer distribution patterns, the abundance development of several, paleobathymetrically significant benthonic species (Tab. 6-1) were compared with each other and with the P/B ratio. The few trends which can be observed do not contradict the paleobathymetric model in general. For example, the similar distribution patterns of St. excolata and D. oxycona in section F, and St. excolata and N. semireticulata in section A7 could be explained by the fact that all three species are characteristic for the outer neritic / upper bathyal and deeper environments (Saint-Marc, 1992; Van Morkhoven et al., 1986; Berggren & Aubert, 1975). The general inverse relationship in distribution between St. excolata and Sp. kneheli, matches the paleobathymetric model as well, because Sp. kneheli reaches its highest abundances in shallower inner neritic environments (Saint-Marc, 1992). The same applies to the distribution of N. semireticulata. Highest abundances are often associated with high planktonic values which corresponds to its relatively deep preferences. Especially remarkable is the co-occurrence of higher abundances of D. oxycona and planktonic peaks in the high-frequency P/B cycle of section F. The latter relationship could be easily (mis-) interpreted as orbitally induced high frequency sea-level changes. However, we assume that all the observed trends in foraminiferal distribution are rather complex results of changes in paleoproductivity than in sea level for various reasons:

a) Quite a number of paleobathymetrically unexpected relationships are developed. For example, an increase in the abundance of the upper bathyal form D. oxycona can often be observed towards lower planktonic values. Highest abundance of the shallower form Sp. kneheli in section P is associated with high planktonic values.

b) Microfossil distribution patterns seem to be clearly dominated by paleoproductivity. Therefore, it seems rather unlikely that any detailed sea-level signal can be detected within these patterns. The striking regional differences in modulation of the regional low-frequency P/B pattern with significant variations in amplitude and frequency can also be best explained by complex paleoproductivity patterns rather than by more unidirectional basinwide sea-level changes.

c) Paleoproductivity related shifts in the depth of the oxygen minimum zone (e.g. Almogi-Labin et al., 1993) are capable of causing changes in the benthic fauna which may have a similar character as paleobathymetrically induced faunal changes.

d) No description of significant eustatic sea-level changes of 4th order or higher from the Late Maastrichtian is known to us (e.g. Haq et al. 1987). Although a climato-eustatic
coupling for orbitally induced high frequency sea-level changes has been described for the Cretaceous (Jacobs & Sahagian, 1993), the associated sea-level amplitudes of several meters should be too small to be detected in the hemipelagic (see also Queene & Ferry, 1995; Glancey et al., 1993; Strasser, 1994; Batt, 1996). Since part of the sections are located on the tectonically quiet shelf (P, F), tectonically induced high frequency sea-level changes (e.g. Malarte & Ferry, 1995) are not expected for the southern sections. Although cyclicity of benthonic abundances does not reflect paleobathymetric changes, the average benthic faunal composition with a typical Midway Type Fauna (50-200 m; Berggren & Aubert, 1975) and some elements of the Velasco Type Fauna (>200 m; Berggren & Aubert, 1975; e.g. D. oxycona) (Tab. 6-1) suggest an outer neritic / upper bathyal depositional environment for the successions. However, the general absence of the bathyal-abyssal marker species Nuttallides truempyi (Nuttall) (e.g. Van Morkhoven et al., 1986) with its rare occurrence in only a few samples (upper thirds of sections A7 and P, respectively), seems to point to a facies not deeper than upper bathyal. The absence may be alternatively explained by the oligotrophic character of this species (Wedmark & Speijer, in press). The outer neritic / upper bathyal model for the Late Maastrichtian of Sinai matches the paleobathymetric interpretation of previous works in the region (e.g. Shahin, 1990; Almogi-Labin et al., 1990a; Abbass et al., 1994; Hewaidy & Ashwah, 1993) as well as the paleogeographic reconstructions for the area by Camoin et al. (1993) and Kuss & Bachmann (1996). It should be noted that the present investigations did not include the approximately 5 m thick hard calcareous marl bed at the top of the Maastrichtian (sections C, F, P, Fig. 6-2), which is interpreted to be associated with a period of shallowing and reduced sedimentation, as has already been shown in a previous study (Lüning et al., submitted b).

The study suggests that the studied foraminiferal fauna of the A. mayaroensis Zone in Eastern Sinai cannot be used for detailed, high resolution paleobathymetric reconstructions. Neither the P/B signal, nor the abundance distribution of the investigated benthonic foraminiferal marker species seem to help in estimating the exact regional sea-level development of that period, because the paleoproductivity component seems to play a more important role here than previously expected for this interval. Therefore, detailed paleobathymetric models for the Late Maastrichtian of Sinai which were predominantly based on foraminiferal data (e.g. Abbass et al., 1994; Shahin, 1990; Hewaidy & Ashwah, 1993) may need to be re-examined.

Comparison with the development in the Paleocene

The strong paleoproductivity control on the microfossil abundances in the Late Maastrichtian contrasts with the results from the Paleocene of Central East Sinai, where it was possible to show that the P/B pattern can be used as a reliable proxy for sea-level changes (Lüning et al., submitted b). The regional switch from productivity to paleobathymetry as a major control mechanism on foraminiferal distribution might be related to the termination of an upwelling system, which according to Almogi-Labin et al. (1993) lasted from the Santonian to Late Maastrichtian. A higher productivity in the Late Maastrichtian relative to the Danian was already described by Benjamini et al. (1990) from Israel and Widmark & Malmgren (1992) from the South Atlantic. Furthermore, it is important to note that during the Late Maastrichtian just before the K/T biotic crisis, the global deep water masses are interpreted to have experienced a significant cooling (Widmark, 1995) which may be connected to the oceanographic change here described for Eastern Sinai.

6.8. Conclusions

Detailed investigations of calcareous nannofossils, planktonic and benthonic foraminifera as well as sedimentological observations in four surface sections were used to evaluate the paleoceanographic and climatic conditions in the Late Maastrichtian intrashelf basin of Eastern Sinai. The Late Maastrichtian Ahathomphalus mayaroensis Zone in Eastern Sinai can be identified easily by the occurrence of the zonal fossil A. mayaroensis. The zone is biostratigraphically further partitioned by the first occurrences of Plummicrita reicheli (ex P. hantkeninoides) and Micula prinsii. A higher resolving cyclostratigraphic subdivision of the interval is proposed based on a low-frequency cyclicity in the P/B ratio. Microfossil abundances and lithologies are characterized by pronounced repetitive distribution patterns. This includes low- and high-frequency P/B fluctuations, repetitive changes in the abundance of calcareous nannofossils and benthonic foraminifera, as well as the development of chalk marl couplets and thinning-upward chalk bundles. Both microfossil distribution patterns and rhythms are attributed to changes in paleoproductivity. Semiquantitative investigations of calcareous nannofossils and benthonic foraminifera allowed to differentiate a high (HP) and low (LP) productivity assemblage. While the HP assemblage is dominated by Glaukolithus diplogrammus, Manivitella pennoatae, Microrbhodus decoratissimus and Micula mura and the benthonic foraminifera Neogloboquadrina pachyderma, the LP assemblage is characterized by Lithraphidites quadratics and Bolivinoides draco. For the chalk marl couplets, thinning-upward chalk bundles and the high-frequency P/B pattern, productivity changes are interpreted to have been related to orbital forcing. High resolution reconstructions of the Late Maastrichtian sea-level history from foraminiferal abundances do not seem useful for Eastern Sinai, mainly due to the strong paleoproductivity signal in the microfossil assemblages. The studied hemipelagites have obviously been deposited during the latest phase of the southern Tethyan upwelling system, which was active from the Santonian to the Late Maastrichtian with a peak in the Campanian (Almogi-
Labin et al., 1993). Termination of upwelling just before the K/T boundary provides a good explanation for the change towards a paleobathymetric control on foraminiferal distribution as observed for the Paleocene of Central East Sinai (cf. Lüning et al., submitted b).

Warning: The area around Gebel Misheiti is thought to contain uncleared landmines from the Sinai wars. During the survey, a large tank mine was found on the desert floor which may mean that more of the explosives are undiscovered in the soil. Gebel Misheiti can best be approached via an asphaltic road starting just west of Themed leading towards the south. The last km has to be done offroad preferably using the tracks of a previous vehicle.
Abstract

Tracking sequences in the hemipelagic facies is traditionally considered as complicated because of the rather monotonous appearance of the hemipelagic deposits, lack of subaerially exposure and missing typical sequence stratigraphic geometries. This study tries to contribute to the understanding of how the sequence stratigraphic model can be applied in hemipelagic environments. Eight latest Maastrichtian to early Eocene hemipelagic sections from Central East Sinai have been recorded and biostratigraphically and paleobathymetrically studied by means of planktonic and benthonic foraminifera, calcareous nannofossils and sedimentological observations. Sea level changes were reconstructed using the foraminiferal planktonic/benthonic ratio, benthonic foraminifera, hiatuses, and hard calcareous marl beds. Correlation of the observed paleobathymetric cycles yielded a highly consistent sea level history for the region with the existing literature taken into account. Candidate sequence stratigraphic surfaces were interpreted using a simple model which provides an approximate relationship between relative sea level and systems tracts. Sequence stratigraphic history was compared with regional sea level curves from Central- /Southern Egypt, Tunisia and Texas, as well as with the European (Hardenbol et al., in press) and the ‘global’ (Haq et al., 1987) sequence charts. In general, an excellent correlation is observed in all cases for at least the early and middle Paleocene (K/T to mid Gl pseudomenardii Zone). Deposition in Central Sinai during this time may therefore have been predominantly controlled by eustatic sea level rather than local tectonics. Sea level development in the late Thanetian is more ambiguous and is discussed. The study demonstrates the applicability of the sequence stratigraphic concept in suitable hemipelagic environments. The Paleocene hemipelagites of Sinai obviously represent a paleobathymetric interval suitable for sea level changes to be recorded in the foraminiferal faunal composition in terms of the foraminiferal planktonic/benthonic ratio.

7.1. Introduction

Sequence stratigraphy was originally developed for passive siliciclastic shelf margins emphasizing temporal-spatial changes in downslope transportation and accumulation in relation to changes in sea level, sediment input and subsidence (Vail et al., 1977). Later, the concept was more or less successfully extended to be used in other environments such as carbonate platforms (e.g. Handford & Loucks, 1993) and fluvial systems (e.g. Wescott, 1993). The relatively shallower marine sedimentary deposits, for which the sequence stratigraphic concept was initially defined, commonly exhibit lithologic variabilities and stratigraphic geometries suitable for sequence stratigraphic interpretations, whereas hemipelagites usually lack features like onlap, downlap, omission surfaces and cyclic lithofacies changes due to monotonous continuous sedimentation, making application of sequence stratigraphic concepts in hemipelagic settings more problematic. Furthermore, hiatuses in hemipelagic environments are often formed by strong ocean floor currents and may be unrelated to sea level changes (Moore et al., 1978). Systems tracts cannot be distinguished in hemipelagic settings by conventional methods. On the other hand, basinial deposits potentially provide better biostratigraphic data, especially from planktonic foraminifera and nannofossils. This is because of more favourable conditions for living and for preservation in deeper water environments. In addition, paleobathymetric effects of varying sedimentary supply can be neglected in hemipelagic environments (see Hancock, 1989). In light of basinwide reconstructions and modelling, however, a correlation between the two environments using a sequence stratigraphic frame is obviously needed. The present study tries to bridge that gap by interpreting paleobathymetric data derived from hemipelagial foraminiferal deposits under a sequence stratigraphic view. Other approaches towards sequence stratigraphy in hemipelagic settings were undertaken by Hancock (1989), Gale (1996), and Ernst et al. (1996) who studied the Upper Cretaceous chalk of Western Europe. They used a number of criteria for the recognition of systems tract boundaries in chalk environments including hardgrounds. Except in a few stratigraphic horizons, hardgrounds or comparable deposits are not developed in the hemipelagites of Central Sinai, probably because the high content of clay minerals prevent carbonate lithification. Robaszynski et al. (1993) presented a study from a ‘distal environment’ in the Cenomanian of Central Tunisia, using also several parameters for the sequence stratigraphic interpretation. However, the facies studied in Tunisia
were still influenced by platform-derived sediment fluxes in contrast to the rather pure hemipelagic conditions in Central Sinai. In addition, Robaszynski et al. (1993) state that most of the systems tracts can be discriminated in the field which is not the case in the present study on Sinai. Armentrout et al. (1990, 1993) and Armentrout (1996) used microfossil abundance patterns and other physico-chemical parameters to locate condensed sections and sequence boundaries in the Plio-Pleistocene of the Gulf of Mexico. A basic assumption in Armentrout’s models is that total microfossil abundance patterns reflect changes in sediment accumulation rates which are interpreted to be associated with sea level changes in terms of condensation. Applicability of this model for Sinai is discussed later in the text. A reconstruction of eustatically controlled systems tracts in the Lower Cretaceous of SE France has been presented by Magniez-Jannin & Jacquin (1990) and Magniez-Jannin (1992) on the basis of the analysis of bentonic foraminiferal assemblages.

7.2. Regional setting

During the Paleocene, Sinai was part of the broad northern shelf margin of the Afroarabian Plate. The Gulf of Suez and Gulf of Aqaba, which bound the Sinai microplate today, were still closed. While Northern Sinai and the adjacent Negev were affected by transpressive movements since the Turonian onwards, the study area in Central and Southern Sinai is thought to have remained tectonically rather calm throughout the whole Mesoozoic and Early Tertiary (Said, 1962; Cohen et al., 1990; Kerdany & Cherif, 1990). The SE-vergent, NE-SW trending domal anticlines in Northern Sinai are part of the ‘Syrian Arc’ (Krenkel, 1924, 1925) which represents an intraplate foldbelt extending from Egypt to Syria formed by Late Cretaceous to recent inversion of Late Triassic / Liassic halfgrabens (Moustafa & Khalil, 1990; Chaimov et al., 1992; Shahar, 1994). Following the nomenclature of Said (1962) the inversion zone can also be called ‘unstable shelf’, and the southern tectonically calm block (including the study area) ‘stable shelf’ (map in Fig. 7-1). Nevertheless, it cannot be fully excluded that tectonic processes acting on the unstable shelf may occasionally have influenced the stable shelf. Because the compressional movements are oriented to southern directions, loading and relaxation effects in terms of a foreland basin model may not be completely ruled out.

At Paleocene times, Central Sinai was part of a northeast-southwest striking intrashelf basin that was confined on the south by the Afroarabian paleoshoreline in the area of the modern Red Sea (about 300 km south of the Sinai Peninsula), and on the north by a number of exposed domal anticlines of the NE-SW striking foldbelt (paleogeographic, paleobathymetric and isopach maps in Bartov & Steinitz, 1977: 142; Sestini, 1984: 169; Said, 1990: 469; Dercourt et al., 1993; Speijer, 1994a: 123). Various studies of foraminifera showed that throughout the Paleocene, a middle to outer shelf environment prevailed in Central Sinai (Hewaidy, 1987; Shahin, 1990, 1992; Ismail, 1992; Speijer, 1994a: 123; Abbass et al., 1994). The hemipelagic sedimentation was dominated by commonly bioturbated, greenish chalky foraminiferal marls which are classified as ‘Esna Shale Formation’ in Northern and Eastern Sinai (Said, 1990: 454; Ziko et al., 1993). Because the region is interpreted as having not been affected by major tectonic processes during the Paleocene, resulting in a uniform subsidence, and hemipelagic sedimentation rates were relatively low, changes in paleodepth may be attributed solely to eustatic sea level changes. Paleobathymetric variations resulting from major vertical sea floor movements and prograding sedimentary bodies may be excluded. Further evidence for the eustatic nature of the sea level cycles, for example correlation with age-equivalent cycles from other basins, will be presented later in this contribution.

7.3. Methods

While previous investigations on the Paleocene of Central Sinai concentrated on rather few isolated sections (mainly in West Sinai: Abbass et al., 1994; Anan, 1992; Ayyad & Hamama, 1991; Marzouk & Abou-El-Enein, 1995; Shahin, 1990) this contribution presents a regional synthesis tying literature data into a newly recorded dense biostratigraphic and paleoecologic framework. During two field expeditions in the Sinai in 1995 and 1996, eight sections from the Paleocene marls were measured on the stable shelf and were sampled at intervals ranging...
Legend

Sections:
- Foraminifera
- Calcareous
- Nannofossils
- Plankton
- Biozones
- Lithologic column

Lithologic symbols:
- Limestone
- Calcareous
- Shelly chalk
- Shelly marl
- Foraminifera
- Biozones
- Isolated biozone
- Biozone boundaries
- Cycle boundaries
- Shale appearance

Biozone boundaries:
- Exact biozone boundary
- Approximate boundary or biozone data missing beyond this boundary
- hiatus / unconformity

Cycle boundaries:
- Direct evidence for cycle boundary (sea level drop)
- Missing evidence for cycle boundary (erosion or too large sample spacing); cycle boundary interpolated from neighbouring sections

Section C near Sheikh Attiya

Section M Taba

Section T1 N'Nuweiba

(base map modified after Jenkins 1990, tectonic map after Neev 1975 and Agah 1981)
between 1-3 metres (Fig. 7-1). To isolate the foraminifera from the matrix, the samples were washed twice through a 63μm sieve after treatment with H₂O₂ and the highly concentrated tenside REWOQUAT, respectively. The microfossil residue was then fractionated into four grain size classes for easier handling. The planktonic foraminifera were determined under the light microscope (by Lüning) following the species- and zonal concepts of Toumarkine & Luterbacher (1985), Blay (1969), and Berggren & Van Couvering (1974). Determination and paleobathymetrical interpretation of the benthonic foraminifera were done (by Lüning) following criteria of Aubert & Berggren (1976), Berggren & Aubert (1975), Luger (1985), LeRoy (1953), Said & Kennaway (1956), Saint-Marc (1992), Speijer (1994a and written communication, 1996) and Van Morkhoven et al. (1986).

For the calcareous nanoplankton, smear slides were prepared using techniques described in Bramlette & Sullivan (1961) and Hay (1961, 1965). The slides were examined (by Marzouk) under the light microscope at a magnification of about x1250 by both cross-polarized and phase-contrast. Biostratigraphy is based on the zonal concept of Martini (1971).

7.4. Paleodepth indicators

7.4.1. Planktonic/Benthonic Foraminiferal Ratio

The foraminiferal planktonic/benthonic (P/B) ratio was routinely determined for all samples counting traverses with a minimum of 300 specimens in the unfractionated washed residue. The value is calculated in percent plankton or benthos in relation to the total number of counted foraminifera.

The P/B ratio serves here as a general paleobathymetric indicator. Under normal conditions, the percentage of planktonic foraminifera in foraminiferal bottom assemblages increases with increasing water depth. The ratio is considered to depend on the relative difference between the productivity of planktonic species, which are in greater densities in open oceanic environments, and the productivity of benthonic foraminifera which is higher in neritic environments than in the deeper oceanic ones (Phleger, 1964; Reiss et al., 1974; Gibson, 1989; Van der Zwaan et al., 1990). However, because productivity of both foraminiferal groups is controlled by numerous water mass properties such as temperature, pressure, density, nutrients, salinity, light penetration, oxygen and other physical, chemical and biological factors, the variations of planktonic percentages for a given depth can be considerable. Therefore, the P/B-ratio provides no numerical depth information but should be used as an indicator for relative paleobathymetric changes. Attention must be paid to depth-independent planktonic abundance events which might be controlled by local climatic and oceanographic effects such as upwelling, oxygen depletion events, storms, or foraminiferal evolutionary trends. Detailed discussions of the distribution of planktonic and benthonic foraminifera are found in Be (1977), Boltovskoy & Wright (1976), Murray (1973, 1991), Vincent & Berger (1981), Gibson (1989), and Van der Zwaan (1990).

Because of the multitude of parameters and processes capable of disturbing the P/B - depth relationship, the P/B-curves in this study are treated with considerable caution. Confirmation of the paleobathymetric significance is sought by intra- and inter-basinal correlations, the latter based mainly on literature data which often take into account paleobathymetric indicators other than foraminifera [e.g. seismic stratal patterns in Haq et al., 1987; see also comparisons with other studies below]. Furthermore, P/B patterns are only employed as paleobathymetric proxies in intervals, where marine paleoproductivity is assumed to have been relatively low.

7.4.2. Benthonic Foraminifera as Paleodepth Indicators

Many species and genera of benthonic foraminifera are characterized by distinct broad depth ranges (for the Paleocene see Berggren & Aubert, 1975; Aubert & Berggren, 1976; Van Morkhoven, et al., 1986; Saint-Marc, 1992; Speijer, 1994a; Luger 1985). As discussed earlier, a multitude of physico-chemical parameters limit the distribution of benthonic foraminifera. However, because these parameters often vary predictably along a depth gradient, benthonic foraminifera can be used for approximate paleobathymetric trend analyses (Murray, 1991; Speijer 1994a,b). Again, only broad depth ranges and trends can be reconstructed rather than exact metrically specified depth intervals.

7.5. Sequence stratigraphic concept

Sea level reconstructions in this study are mainly based on microfossil data supported by sedimentological observations so that the method used might be also described by the term 'sequence biostratigraphy' sensu Armentrout (1996). Biostratigraphically well constrained, longterm, hemipelagic cycles are interpreted as sequences in this contribution, if there is evidence that the cycles have been largely controlled by sea level changes. As will be shown in detail, the P/B-ratio is considered here as a paleobathymetric proxy, allowing the reconstruction of sea level cycles and their relative amplitudes, although, only within a non-quantitative frame. Geometry of the P/B-curve (hinge and inflection points, positive / negative / constant trends) is used to estimate
the approximate position of candidates for the transgressive surface (ts), the maximum flooding surface (mfs) and the sequence boundary (sb) (Fig. 7-2). The position of the three sequence stratigraphic surfaces in the curve follows concepts established for eustatic sea level curves used by many authors (e.g. Handford & Loucks, 1993). In our study, the ts is placed just above the first significant increase in planktonic foraminiferal values, the mfs just below the planktonic maximum and the sb above the first major drop in planktonic values (Fig. 7-2). Correlation with similar P/B-patterns and other parameters of other sections within a chronostratigraphic framework is done to evaluate whether or not a candidate surface has a regional distribution, and by, that represents a sequence stratigraphic surface. One basic assumption is that the P/B-trends co-vary with the general stacking patterns developed in shallower regions of the basin.

Similar principles for systems tracts differentiation based on curve geometry are applied for wireline logs (e.g. Emery & Myers, 1996) as well as for carbon isotope curves in shallow marine series (Vahrenkamp, 1996: Tab. 1) and in chalks (Mitchell et al., 1996). For a summary of the application of micropaleontology in sequence stratigraphy see Simmons & Williams (1992).

Because the systems tracts boundary surfaces in the hemipelagic do not represent physical boundaries but ‘correlative conformities’ (Vail et al., 1977), only intervals rather than discrete surfaces can be reconstructed for the position of the sequence stratigraphic surfaces. However, attributing biostratigraphic ages to potential systems tracts boundaries in the hemipelagic allows correlation of the age information up-dip towards the shallow marine strata where biostratigraphic data are often lacking but boundaries are more clearly developed. As a precondition, of course, there must be evidence that the shallow marine sequences are identical to the hemipelagic cycles, substantiated for example by similar stacking patterns and corresponding number of cycles.

Shifts in sequence boundary ages of up to one-half of a sea level cycle are postulated by Steckler et al. (1993), which contributes to the overall uncertainty of the determination of the boundary surface age. Although sequence geometry is most sensitive to sea level changes, other factors, including the subsidence rate, sediment supply, sediment loading and compaction, affect the timing in systems tracts distribution. In complex basins, therefore, timing of the transgressive surface, maximum flooding surface and sequence boundary might vary laterally. Consequently, indicating intervals alone, rather than exact physical boundary surfaces in sections of hemipelagic deposits, may not represent a significant problem if compared to the uncertainties associated with lateral changes in sequence boundary timing (Steckler et al., 1993) and the limited biostratigraphic resolution.

7.6. Paleocene sequences in central Sinai

7.6.1. Biostratigraphic Resolution

The Paleocene sections from the stable shelf of Central Sinai studied here are characterized by mostly excellent micro- and nannofossil preservation and an overall high degree of stratigraphical completeness. Though hiatuses are only minor they are nevertheless significant. Dual biozonation by planktonic foraminifera and nannofossils leads to a high-resolution biostratigraphic frame in which one fossil group may compensate for the occasional weak time resolution of the second group. For example, the Gl. pseudomenardii Zone consists of four nannofossil-zones and the nannofossil-zone NP4 of three planktonic foram zones. The correlation between planktonic foraminifera- and nannofossil-zones is consistent for all sections from Sinai examined here (Fig. 7-3), but in some parts differs markedly from other published schemes such as in Berggren et al. (1995) or Bolli et al. (1985). Possible reasons for the observed discrepancies include lateral diachrony, preservational problems, undeciphered unconformities and differences in taxonomic concepts etc. (discussed in detail for eastern Sinai by Marzouk & Lüning (submitted)). As an example for the data base, the biostratigraphic data set for section C (Sheikh Attiya, see Fig. 7-1) is illustrated in Figure 7-4.
7.6.2. Paleoceanographic and Paleobathymetric Situation

The early and middle Paleocene of Northern Africa is characterized by low marine productivity which provides an important precondition for the paleobathymetric interpretation of P/B patterns in Sinai. The phase of low productivity started with a regional switch from high to low surface water productivity around the K/T boundary which has been described from Israel (Benjamini et al., 1990; Almogi-Labin et al., 1993), Egypt (Lünstig et al., submitted c), Tunisia (Keller, 1988; Keller & Lindinger, 1989; Kouwenhoven et al., 1997), and the northeast African region (Speijer & Van der Zwaan, 1996). Towards the end of this low productivity phase, upwelling in Egypt is interpreted to have gradually intensified during the latest Paleocene (Speijer et al., 1996). From the latest Paleocene of Tunisia, a possibly even more pronounced eutrophication with accumulation of phosphoritic sediments (Aubert & Berggren, 1976; Salaj, 1980, 1986) is reported (Donze et al., 1982; Peyroutet et al., 1986; Kouwenhoven et al., 1997). The paleobathymetric reconstructions in this study, therefore, mainly focus on the low productivity period and do not include the M. velascoiensis Zone of the latest Paleocene. Our sea level interpretations for the Gl. pseudomenardii Zone in part already involve sediments which have been accumulated under gradually increasing paleoproductivity conditions (Speijer et al., 1996). The influence of productivity on these paleobathymetric reconstructions and their basis is discussed below. Exceptional high frequency changes or blooms in foraminiferal and nannofossil assemblages, which are indicative of a major productivity control (e.g. in the Late Cretaceous: Eshet & Almogi-Labin, 1996; Lünstig et al., submitted c), have not been observed in any interval of the Paleocene sections.

Besides the P/B patterns, paleobathymetry in this study is based on the distribution of hiatuses, hard calcareous marl beds, and the analysis of benthonic foraminifera. Hiatuses will only be recognized here if a foraminifera- or nannofossil-zone is missing. It must be mentioned that the present study is focused on vertical sea level changes rather than horizontal movements of the coastline (transgressions, regressions). Although in most cases a sea level rise may be associated with a transgressive phase, this need not always be the case, e.g. in areas with regional tectonic tilt (see Hancock 1989, and Lewy 1990). The terms 'transgression' and 'regression' are, therefore, intentionally avoided in this contribution. The following examples are to summarize how the depth tools are used and which relationships exist between them.

The sections yielded P/B-curves with values covering the whole spectrum between 100% and 0% plankton (P). A number of typical, well-correlatable benthos peaks with significantly lowered plankton percentages is clearly developed and can be traced through many sections. For example, there is a pronounced composite benthos peak in the Gl. pseudomenardii Zone (NP5-NP9) which is characterized by a complex interval of very low P-values. A second prominent benthos peak is developed in the M. angulata Zone (mid NP4). In some sections, these two peaks are replaced by or associated with significant hiatuses, which probably indicate a lowered base level associated with increased winnowing currents (e.g. in sects. K, P, R; Fig. 7-1) (see also Bathurst, 1975, and Brett, 1995). The occurrence of hiatus development may be confined to low relief submarine swell regions.

Indurated beds of late Paleocene calcareous marl and marly chalks are interpreted here to have formed during periods of lowered sea level, because they are sandwiched between softer marls with low planktonic values (e.g. K, P, Q, T 1). In one section it was possible to confirm this interpretation by correlating the hard bed with a well-developed benthonic peak (section R, cycle ThSin-5/YpSin-1, Figs. 7-1 and 7-3). In the field it seems that the hard beds consist of the same material as the softer chalky marls below and above and are probably distinguished mainly by a lower porosity and by calcitic cement. It seems reasonable that they might represent firmgrounds or immature hardgrounds similar to those described in the Chalk of Northwest Europe (Bathurst, 1975; Bromley, 1965; Hancock, 1989; Gale, 1996). In analogy to Northwest Europe we assume that during sea level fall, bottom currents might have accelerated due to the reduced space for the water exchange. These slightly stronger currents might have been capable of winnowing and even eroding the sediment (Hancock, 1989; Brett, 1995: 603). Winnowing might have exposed the upper layers of the sediment surface to an increased pore water exchange with the sea water for a considerably long time, leading to a pronounced cementation and lithification (e.g. Allouc, 1990). Thicker, amalgamated hardbed units of this kind may be the result of repeated phases of non-deposition, lithification and sedimentation (Bathurst, 1975; Bromley, 1965, 1975; Gale, 1996). Another possible mechanism of producing hard beds sandwiched between softer hemipelagics during phases of lower sea level is linked to the change of aragonite solubility with depth. Whereas during times of high sea level, the metastable aragonite is mainly dissolved and carried away, aragonite is more stable in times of reduced water depth and may be available for cementation during early diagenesis (Bathurst, 1975; Hudson, 1967).

Figure 7-3 (following page).Paleocene relative sea level history and sequences of Central East Sinai as reconstructed in this study (no scale implied). Correlation with a sequence chart for Europe (Hardenbol et al., in press) and a 'global' chart (Haq et al., 1987) is attempted in the last two columns at the right. In general, the Paleocean sequences show a good degree of correspondence, at least until the mid Thanetian (see text for further details). Note the good correlation between Hardenbol et al. (in press) and this study in the Gl. pseudomenardii Zone in terms of timing and number of sequences whereas Haq et al. (1988) postulated a pronounced sea level rise already in NP7. Correlation between planktonic foraminifera- and nannofossil-zones is consistent for all studied sections from Sinai but differs sometimes markedly from other published schemes like in Berggren et al. (1995) or Bolli et al. (1985).
In a few cases benthonic foraminifera were studied semi-quantitatively in order to test the paleodepth changes that were reconstructed with the help of P/B-data. Paleobathymetric information was gained by examining changes in the abundance of deep and shallow water indicator species rather than appearance or non-appearance of these species (sections C-R, cycles ThSin1/2 and ThSin2/3) (see discussion at ThSin1+2).

Armentrout et al. (1991) and Armentrout (1996) used microfossil abundance patterns of several microfossil groups in Plio-Pleistocene sediments to interpret the sequence stratigraphic history. Following their ideas, higher microfossil abundances may be related to a relatively higher sea level which leads to reduced terrigenous sedimentation and therefore to a fossil concentration during transgressive periods (condensed section at the maximum flooding surface, e.g. Loutit et al. 1988). Sediments deposited during shallower periods may be characterized by decreased microfossil abundances caused by depositional dilution and increased environmental stress. In the Sinai study, the microfossil abundance was quantified with respect to the foraminiferal P/B-ratio only, because many of the parameters mentioned by Armentrout can be interpreted in a number of ways and investigations of group abundances are very time consuming in relation to their paleobathymetric significance. Nevertheless, the trend of increasing total microfossil abundance towards deeper waters can be confirmed with the material studied from Sinai. Microfossil abundances clearly correlate positively with higher planktonic foraminifera percentages, as simple qualitative observations of the microfossil washing residue showed.

Interpretation of the relative amplitudes and the stacking patterns of the different sea level cycles (Fig. 7-3) is mainly based on P/B data in combination with the sedimentological models for the hiatuses and firm-hardgrounds discussed above.

7.6.3. Detailed Description and Correlation of the 3rd Order Sequences Studied

Within the the uppermost Amastraichtian to lower Eocene interval, ten 3rd order sequence boundaries were reconstructed (Figs. 7-1 and 7-3). The 3rd order character is based on the typical cycle durations which vary between 0.3 to 3 My. The sequence boundaries are numbered following a system used by Hardenbol et al. (in press). The first two letters refer to the stage (e.g. Th for Thaestian), and the letters 'Sin' indicate that the sequence boundary was reconstructed for the Sinai Peninsula and neighbouring areas. The letters are followed by a running number counted separately for each stage. Sequences are named after the lower and upper cycle boundary. Succeeding transgressive surfaces (ts) and maximum flooding surfaces (mfs) are in most cases interpreted based on the first rise in the P/B-curve (ts) and near maximum plankton values (mfs) according to the scheme in Fig. 7-2. A complete compilation of boundary ages, explanations and references is summarized in Table 7-1. The sections are illustrated in Fig. 7-1. The descriptions below are listed from the oldest to the youngest cycle boundary.

MaSin-Z

Interpretation of cycle boundary MaSin-Z (late A. mayaroensis Zone / nannofossils; M. prinsii Zone) is based on the presence of a regionally characteristic, 5 m thick, hard bed of chalky calcareous marl which becomes relatively softer towards the top. This bed is directly underlain by a chalk with 65-95 % planktonic foraminifera. A thin section of the lower part of the hard bed contained planktonic foraminifera but no globotruncanids, indicating deeper water (Hart 1980), which, in contrast, are highly abundant in the underlying chalk. The bed contains a number of reddish-black/yellowish scattered ferruginous concretions (several cm in diameter) probably associated with phases of reduced sedimentation. In addition, separate clusters of branched, tubular (diam. 1-2 mm) trace fossils with a reddish-brown ferruginous burrow infill are found. The tubes exhibit circular cross sections and are therefore not compacted, which might be regarded as evidence for early lithification of the hard bed. The hard bed may be of similar origin as the upper Paleocene hard beds discussed earlier in the text. A lowered sea level during the late A. mayaroensis Zone may have led to increased velocity and turbulence of bottom flowing currents, consequently leading to reduced sedimentation and increased cementation of the uppermost sediment layers. The bed may also be attributed to a winnowing transgressive phase (Gale. 1996). In the latter case, however, a preceding phase of low sea level is again required, since plankton values below and in the upper part of the hard bed are similar, thus indicating comparable paleobathymetric conditions. Because of the absence of several typical hardground indicators such as borings, sessil or encrusting organisms etc., the bed might be interpreted as a firmground or immature hardground. A more detailed investigation of this pre-K/T-bed in the future would obviously be of great value. The postulated sea level low is also documented as a pronounced P/B-drop in SW-Sinai (Wadi Feiran, Shahin, 1992). Comparable latest Amastraichtian sea-level lowstands were described from numerous sections globally, e.g. Denmark (Schmitz et al., 1992).
<table>
<thead>
<tr>
<th>3rd order sequences (sequence boundaries numbered)</th>
<th>Systems tracts boundaries</th>
<th>Biozonal boundary ages</th>
<th>Basis for systems tract boundary interpretations</th>
<th>Evidence for validity of the cycle boundaries as provided by other authors</th>
</tr>
</thead>
<tbody>
<tr>
<td>YpSin-1</td>
<td>sb M. subbotinae or M. formosa-formosa</td>
<td>NP11 or early NP12</td>
<td>sb at base of hard calcareous foraminiferal marl</td>
<td>Limestones of the Thebes Formation in SW-Sinai interpreted as deposited in open marine inner neritic conditions in contrast to underlying middle neritic to bathyal marl succession (Shahin, 1990)</td>
</tr>
<tr>
<td>mfs early M. edgar</td>
<td>around NP9 / NP10 boundary</td>
<td>HST-interpretation based on high P/B values</td>
<td>high P/B values in the Egyptian Western Desert (Luger, 1985), transgressive period in the NE-surroundings of the Gulf of Suez based on foraminiferal and sedimentological evidence (Kulbrok, 1996)</td>
<td></td>
</tr>
<tr>
<td>ts ?</td>
<td>?</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ThSin-5 sb late Gi. pseudomenardii</td>
<td>around NP7/8 / NP9 boundary</td>
<td>sb at base of hard bed of calcareous marl (probably consisting of 100% B, firmground? see text) (LST-bed developed in sect. K,R); bed replaced by softer marl exhibiting a clear P/B-drop in section K (here sb interpreted after P/B-trend)</td>
<td>hard calcareous marl age-equivalent to upper part of Tarawan Chalk (Gebel Oweina, southern Eastern Desert; fig. 24.3 in Said, 1990)</td>
<td></td>
</tr>
<tr>
<td>mfs Gi. pseudomenardii</td>
<td>late NP7/8</td>
<td>HST-interpretation based on high P/B values greater than zero (e.g. 50% P in sect. R, 10% P in sect. P)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ts Gi. pseudomenardii</td>
<td>NP7/8</td>
<td>ts at top of hard calcareous marl bed</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ThSin-4 sb Gi. pseudomenardii</td>
<td>early NP7/8</td>
<td>HST-interpretation based on high P/B values (cycle developed in sect. C,Q,R)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>mfs Gi. pseudomenardii</td>
<td>early NP7/8</td>
<td>ts HST-interpretation based on high P/B values</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ts Gi. pseudomenardii</td>
<td>late NP6</td>
<td>ts HST-interpretation based on high P/B values</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ThSin-3 sb Gi. pseudomenardii</td>
<td>NP6</td>
<td>sb P/B-drop in sect. C,Q,R, hiatus in sect. T1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>mfs Gi. pseudomenardii</td>
<td>early NP6</td>
<td>ts HST-interpretation based on high P/B values</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ts early Gl. pseudomenardii</td>
<td>latest NP5</td>
<td>ts HST-interpretation based on high P/B values</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ThSin-2 sb around P. pusilla pusilla / Gl. pseudomenardii boundary</td>
<td>mid NP5</td>
<td>very strong P/B-drop from 90% to 10-30% P (e.g. sect. F,Q,T1), locally significant erosional hiatuses (e.g. in sects. K and P cutting down to Pr. uncinata/M. angustata-boundary; this hiatus is probably of multiple origin in combined action with SL-drops at ThSin-1 and ThSin-3)</td>
<td>P/B-drop in Wadi Feiran (Shahin, 1990, 1992), Wadi El-Seig (Abbas et al., 1994) (both Central West Sinai); drop in palaeodepth in Gebel Oweina (Nile Valley, southern Egypt) (based on foraminiferal foraminifera) (Speijer, 1994), hiatus in Abu Zenima area (Central West Sinai, Anan, 1992) and the Western Desert of Egypt (Luger, 1985); sb in NE-surroundings of the Gulf of Suez based on foraminiferal and sedimentological evidence (Kulbrok, 1996)</td>
<td></td>
</tr>
<tr>
<td>mfs</td>
<td>around M. angulata / P. pusilla pusilla boundary</td>
<td>late NP4</td>
<td>interpreted after P/B, relatively fast reestablishment of P/B values as comparable to P/B before cycle boundary ThSin-1</td>
<td></td>
</tr>
<tr>
<td>-----</td>
<td>-----------------------------------------------</td>
<td>---------</td>
<td>-----------------------------------------------------------------------------------------------------------------</td>
<td></td>
</tr>
<tr>
<td>ts</td>
<td>mid M. angulata</td>
<td>NP4</td>
<td>ts interpreted after first rise in P/B curve</td>
<td></td>
</tr>
<tr>
<td>sb</td>
<td>early M. angulata</td>
<td>mid NP4</td>
<td>strong P/B-drop (sect. C,F,M,Q,T1), erosional hiatus in sect. R</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>P/B-drop in Wadi Feiran and Gebel Ekma (Shahin, 1992; Shahin, 1990), Wadi El-Seig (Abbass et al., 1994) (all Central West Sinai); drop in paleodeth in Gebel Oweina (southern Eastern Desert, Egypt) (Speijer, 1994, based on benth. forams.); sb in NE-surroundings of the Gulf of Suez based on foraminiferal and sedimentological evidence (Kulbrok, 1996)</td>
<td></td>
</tr>
<tr>
<td>mfs</td>
<td>mid Pr. uncinata</td>
<td>around NP3/NP4 boundary</td>
<td>HST-interpretation based on high P/B values</td>
<td></td>
</tr>
<tr>
<td>ts</td>
<td>late Pr. trinidadensis</td>
<td>late NP3</td>
<td>ts interpreted after first rise in P/B curve</td>
<td></td>
</tr>
<tr>
<td>sb</td>
<td>Pr. trinidadensis</td>
<td>early NP3</td>
<td>slight long spanning P/B-drop (sect. C,F,M,Q,T1)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Ea. Pa. pseudobulloides</td>
<td>around NP1 / NP2 boundary</td>
<td>HST-interpretation based on high P/B values</td>
<td></td>
</tr>
<tr>
<td>ts</td>
<td>early Pa. pseudobulloides</td>
<td>mid NP1</td>
<td>ts interpreted after first rise in P/B curve</td>
<td></td>
</tr>
<tr>
<td>sb</td>
<td>early Pa. pseudobulloides</td>
<td>NP1</td>
<td>drop in P/B-curve (consistent for sect. C,F,P)</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Si- drop in Hor Hahar (syncline in Negev) (based on calc. nanofoss., dinoflag., forams.) (Eshet et al., 1992)</td>
<td></td>
</tr>
<tr>
<td>mfs</td>
<td>early Pa. pseudobulloides</td>
<td>NP1</td>
<td>HST-interpretation based on high P/B values</td>
<td></td>
</tr>
<tr>
<td>ts</td>
<td>Pa. eugubina</td>
<td>early NP1 around KT-boundary</td>
<td>is-interpretation based on first rise in P/B curve</td>
<td></td>
</tr>
<tr>
<td>sb</td>
<td></td>
<td></td>
<td>sb: reworked Cretaceous plankt., forams contained in marls with high P/B (TST + HST) of NP1 age (sect. C,F,P). Extremely strong SL-drop + subaerial exposure inferred due to lack of shallow marine LST deposits and locally deep reaching erosion (sect. M)</td>
<td></td>
</tr>
<tr>
<td>mfs</td>
<td>latest A. mayaroensis</td>
<td>M. prinsii</td>
<td>HST-interpretation based on high P/B in mod. hard chalky micrite (upper part of 5 m thick micrite bed in sect. C), P/B not documented but should be located in upper part of thick micrite bed (sect. C,F,P), Peak of regressional episode at KT-boundary in Hor Hahar (syncline in Negev) (based on calc. nanofoss., dinoflag., forams., Eshet et al., 1992); very shallow around KT in West Central Sinai (based on plankt. + benth. forams., Ismail, 1992, Abbas et al., 1994); SL-low at KT in Negev (Keller &amp; Benjamini 1991); hiatus at KT in Abu Zenima area (Amann 1992, Marzouk et al. 1994), Gebel Mokattab (Ayad &amp; Hamana, 1991), Wadi Feiran and Gebel Qreiya (Masters, 1984; Marzouk &amp; Abu-El-Enein, 1995), Wadi Feiran section of Shahin (1992); sea level drop during M. prinsii Z. in Central Egypt (Gebel Qreiya, Luger et al., 1997, in press)</td>
<td></td>
</tr>
<tr>
<td>ts</td>
<td>latest A. mayaroensis</td>
<td>M. prinsii</td>
<td>ts interpreted after first rise in P/B curve</td>
<td></td>
</tr>
<tr>
<td>sb</td>
<td>late A. mayaroensis</td>
<td>M. prinsii</td>
<td>sb below, chalky with high P/B (65-89% P) (HST) and an overlying hard bed of 5 m chalky calcite, marl with plankt. forams, but no Globotruncanids (LST) (sect. C,F,P), firmground? / immature hardground? see text</td>
<td></td>
</tr>
</tbody>
</table>

Table 7.1. Timing and sedimentary / faunal evidence of sequences and systems tracts boundaries as found for Central East Sinai during this study. The last column on the right lists references from nearby regional studies which provide data supporting the postulated cycle boundaries presented in this contribution.
The following transgressive surface (ts) and maximum flooding surface (mfs) are expected to lie in the latest *A. mayaroensis* Zone within the upper part of the hard bed as indicated by high plankton values with abundant globostracans in section C. A latest Maastrichtian sea level rise is also reported from Central Egypt (Gebel Qreiya) by Luger et al. (1997, in press) as interpreted from foraminifera of the *K. falsocalcarata* Zone / *M. prinsii* Zone.

**DaSin-1**

The earliest Paleocene (*Pa. eugubina* Zone or *Pa. pseudobulloides* Zone / NPI) is characterized by greenish marls with high plankton values containing reworked Cretaceous foraminifera (e.g., sects, C, P). A similar situation is described from Gebel Qreiya (Central Egypt) by Luger et al. (1997, in press). A significant hiatus was found in the Tabula section (M) and was also recorded by many authors from West Sinai and the Negev (references see Table 7-1). A major sea level fall associated with erosion is assumed. The horizon just above the K/T-boundary is therefore interpreted as a sequence boundary (DaSin-1). The P/B-curve provides no information, because during sea level low non-deposition and erosion prevailed.

The succeeding ts and mfs are interpreted with respect to P/B-values and may be located in the *Pa. eugubina* Zone and early *Pa. pseudobulloides* Zone, respectively (both in NPI).

**DaSin-2**

Cycle boundary DaSin-2 (early *Pa. pseudobulloides* Zone / NPI) is characterized by a benthos-peak which is rather weak but consistent for many sections (sects. C, F, P). An age-equivalent sea level drop is described from the Negev (Hor Hahar, Eshet et al., 1992) as based on paleoecological studies of calcareous nanofossils, dinoflagellates and foraminifera. After the P/B curve, the ts is interpreted to lie in the early *Pa. pseudobulloides* Zone (mid NPI), while the mfs is reconstructed for the NPI/NP2 boundary (within the *Pa. pseudobulloides* Zone).

**DaSin-3**

Interpretation of cycle boundary DaSin-3 (*Pr. trinidadensis* Zone / early NP3) is based on a long-term low-amplitude benthos peak which is rather weak but can be found in a number of sections (sects. C, F, P, Q, T1). Direct evidence for the paleobathymetric significance of this peak is missing, but a comparison with the 'eustatic' curve in Haq et al. (1988) shows a similar sea level low with the same timing and the same long spanning / low amplitude geometry. A sequence boundary in the *Pr. trinidadensis* Zone is also described from the northeastern surroundings of the Gulf of Suez by Kulbrok (1996) as based on foraminiferal and sedimentological data. Based on P/B data, the ts is interpreted to lie in the latest *Pr. trinidadensis* Zone boundary (late NP3), while the mfs is reconstructed to lie within the *Pr. uncinata* Zone (around the NPI/NP4 boundary).

**ThSin-1**

Cycle boundary ThSin-1 is dated as early *M. angulata* Zone / mid NP4 and is developed as a well-resolved, distinct benthos peak in many sections (sects, C, M, Q, T1) and in W-Sinai sections of Shahin (1992) and Abbass et al. (1994). Haq et al. (1988) shows a similar sea level low with the same timing and the same long spanning / low amplitude geometry. A sequence boundary in the *M. angulata* Zone / early NP3) is characterized by greenish marls with high plankton values (around 95%). The cycle boundary is also described from the northeastern surroundings of the Gulf of Suez based on foraminiferal and sedimentological evidence (Kulbrok, 1996). An age-equivalent drop in paleodepth is also reported by Speijer (1994b) from Gebel Owieina in the Nile Valley of southern Egypt based on paleobathymetric investigations of benthonic foraminifera.

Our investigations of benthonic foraminifera along a falling limb of the P/B-curve of section C (decreasing planktonic values) seem to confirm the depth-significance of the P/B-ratio for this interval. All samples investigated yielded a rich assemblage of Midway-type fauna suggesting about 50-200 m water depth (Berggren & Aubert, 1975) as well as some species of the Velasco-type fauna indicating more than 200 m water depth (Berggren & Aubert, 1975). A simple semi-quantitative study showed that the abundance of certain species with deeper water affinity decreases with falling P/B-ratio. This trend was especially observed for *Dorothyia oxycosta* (Reuss) (Velasco-type fauna, >200 m after Berggren & Aubert, 1975; >200 m after Saint-Marc, 1992) and *Neoelliphtella jarvisi* (Cushman) (outer neritic to bathyal). Investigations of ostracodes by M. Morsi (Ain Shams Univ., Cairo, written communication) at the same falling P/B-limb of section C yielded a transition from a rich outer-shelf ostracode fauna to a low abundant and finally an ostracode-free interval. Among the ostracodes of outer shelf affinity, there are *Krithe echolsae* Esker, *Maurisitina coronata* (Esker), *Megommatocthera denticulata* (Esker), *Odonotya ordonii* (Bassiouni) (paleodepth interpretation after Bassiouni & Lugier, 1990). This decrease in ostracode abundance provides further positive evidence for the postulated sea level fall at ThSin-1.
The successive ts is marked by a significant rise in planktonic values (mid *M. angulata* Zone / NP4) while the ms is interpreted near the re-establishment of constantly high values (around the *M. angulata* / *P. pusilla pusilla* boundary/ late NP4).

**ThSin-2**

The most obvious and pronounced sea level drop is developed at the cycle boundary ThSin-2. Around the zonal boundary of *P. pusilla pusilla* and *Gl. pseudomenardii* (mid NP5), the plankton percentages show a strong decrease from as much as 90% to 30-10%. This P/B-drop is also documented in sections from Central West Sinai (Shahin, 1990, 1992; Abbass et al., 1994). The sequence boundary ThSin-2 can be correlated to a sequence boundary interpreted in the northeastern surroundings of the Gulf of Suez (Kulbrok, 1996). The slightly younger boundary age given by Kulbrok (1996) might be explained by the rather poor planktonic foraminiferal fauna in his study area with important zonal marker species missing. Two of the newly studied sections in Central East Sinai and one section from West Sinai (Wadi Feiran area, Anan et al., 1992) contain significant hiatuses at this horizon. Erosion associated with base level fall might have removed all material of the ThSin-1/2-cycle, probably facilitated by a pre-existing hiatus at the ThSin-1 cycle boundary. In sections K and P for example, the *Gl. pseudomenardii* Zone lies directly on top of the *Pr. uncinata* Zone, with two plankton zones missing. In section P, glauconitic marl is developed above the hiatus, probably indicating initial slow rates of sedimentation after the preceding non-depositional or erosional period. Similar hiatuses are described from the Western Desert of Egypt and have been interpreted as being associated with a sea level fall (Luger, 1985).

Studies of benthonic foraminifera by Speijer (1994b) provide evidence for an age-equivalent decrease in paleodepth at Gebel Oweina (Nile Valley of southern Egypt). Our investigations of benthonic foraminifera along a composite falling limb of the P/B-curve in sections C and R (decreasing planktonic values) seem to confirm the depth-significance of the P/B-ratio for this interval (Fig. 7-5). In analogy to the investigations along the cycle boundary ThSin-1 (see above), the three samples studied here contain a rich Midway-type fauna. The only typical representative of the deeper Velasco-type fauna is *Nuttallides truempyi* (Nuttall). In the studied example, this

![Typical elements of the Midway Fauna found in all 3 samples](image)

Only one element of Velasco Fauna found (in all 3 samples): *Nuttallides truempyi* (common in water depths >200 m after Berggren & Aubert, 1975; decrease of abundance with decreasing planktonic values)

![Typical Velasco Fauna elements which were searched for but not found:](image)

*Gavelinella beccariformis* *Bulimina trinitatensis*, etc.

Figure 7-5. Qualitative and simple semi-quantitative analysis of selected benthonic foraminifera in three selected samples across the sequence boundary ThSin-2 (zonal boundary *P. pusilla pusilla* - *Gl. pseudomenardii* / mid NP5) of a composite section compiled from sections C and R (studied samples C1-27, C1-30 and R1-10). General shallowing upward trend as interpreted by falling P/B-ratio is supported by upward decreasing abundance of *Nuttallides truempyi* (Nuttall) [in general deeper than 200 m after Berggren & Aubert (1975) and Van Morkhoven et al. (1986)]. *Spiroplectammina knebeli* Le Roy, characteristic for water depth shallower than 30 m (Saint-Marc, 1992) was only found in the sample with the lowest planktonic percentages (R1-10). All three samples contain a rich Midway-type fauna, typical for water depths between 50-200 m (Berggren & Aubert, 1975).
Cycle boundary ThSin-3 (Gl. pseudomenardii Zone / NP6) is documented by a benthos peak with an intermediate amplitude in three sections (C, Q, R) and by a hiatus contact in section T1. This cycle is part of a composite cycle comprising the cycle boundaries ThSin-2 to ThSin-5 which developed during a relatively long period of low sea level within the Gl. pseudomenardii Zone of Sinai. Discrimination of the individual cycles is only possible in combination with the highly resolving biozonation of calcareous nannofossils. The sediments during this long phase of low sea level were nevertheless deposited far from the coast as documented by bentonic foraminiferal assemblages of Midway-type (see discussion at ThSin-2) and a fossil tooth assemblage with Squaliformes indicative of a deeper shelf environment. The fauna is dominated by sharks with a relatively high diversity (determination of teeth and interpretation by E. Bernadez Rodriguez, Univ. Oviedo, Spain). Skeletal preservation in the sediments deposited during the long period of low sea level is not as good as in the under- and overlying marls. This may be explained by increased currents associated with sediment starvation leading to a longer exposure at the unsheltered sediment water boundary (e.g. Brett, 1995). Other regional studies showed that towards the latest Paleocene, paleoproductivity increases. Nevertheless, the combination of P/B data, hiatuses and hard beds for the paleobathymetric reconstruction of the Gl. pseudomenardii Zone in this study yields a relatively consistent pattern (see discussion at the different sequence boundaries ThSin-2 to 5). In addition, exceptional high frequency blooms in the foraminiferal and nannofossil assemblages, and phosphoritic, cherty or organic-rich sediments were not observed, which excludes the prevalence of pronounced high productivity conditions during this period. It therefore is assumed that productivity may be increased during the time of the Gl. pseudomenardii Zone but sedimentation and the P/B pattern is still dominated by the sea level signal. A different situation with productivity-dominated foraminiferal patterns may be found in shallower segments of the shelf, for example in the southern Eastern Desert of Egypt (Speijer et al., 1996) or in areas which probably are closer to the upwelling centres, e.g. Tunisia (Donze et al., 1982; Peypouquet et al., 1986; Kouwenhoven et al., 1997).

The ts succeeding ThSin-3 is interpreted after the first rise in the P/B curve (around the NP6 / NP7/8 boundary) while the mfs is reconstructed for the early NP7/8 (both within Gl. pseudomenardii Zone).

Cycle boundary ThSin-4 is developed at the base of a hard bed of calcareous marl which lies over marls with higher plankton values. This hard bed was found in many of the sections (C, F, K, P, Q, R, T1) and is of early NP7/8 age. The bed seems to be equivalent to the lower part of the Tarawan Chalk in the Eastern and Western Desert of Egypt [e.g. in Gebel Oweina, Nile Valley of southern Egypt, Fig. 24.3 in Said (1990) with nannofossil zones] based on biostratigraphic age and lithology. We interpret the Tarawan Chalk as being deposited in a shallower water depth than the greenish marls, between which the chalk is sandwiched. A similar interpretation can be found in Anan (1992) where the Tarawan Chalk in the Abu Zenima area (West-Sinai) is considered as having been deposited during a regressional phase.

In a few sections (K, P, R) the hard calcareous marl bed is directly overlain by greenish marl containing 0-50% planktonic foraminifera representing an interval of sea level rise (ts lying within NP7/8, mfs in late NP7/8, both within Gl. pseudomenardii Zone).

Cycle boundary ThSin-5 (around the zonal boundary of NP7/8 / NP9) is documented in only a few sections of the study area. In sections K and P, a second hard calcareous marl bed is developed whereas in section R, this bed is replaced by a clearly resolvable benthos peak. Correlation of the hard bed with a benthos peak gives support to the interpretation that the hard bed is formed during a phase of lowered sea level. In sections C, F and Q, only one hard calcareous marl bed is developed so that the cycles ThSin-4/5 and ThSin-5/6 cannot be discriminated. The amalgamated single calcareous marl beds might be equivalent to the undifferentiated limestones of the Tarawan Chalk unit in the Eastern and Western Desert of Egypt [e.g. in Gebel Oweina, Nile Valley of southern Egypt, Fig. 24.3 in Said (1990) with nannofossil zones; see also Strougo (1986)]. The upper part of the Tarawan Chalk is obviously time-equivalent to the calcareous marl bed above ThSin-5 (second hard bed in sections K, P). Higher plankton values above the calcareous marl bed(s) have been reestablished in the lower M. velascoensis Zone / NP9 (Fig. 7-1) so that a ts is interpreted for this period.
Deposits of the *M. velascoensis* Zone are unusually thin to absent so that reconstruction of the sea level history for this stratigraphic interval cannot be accomplished. Strata may be missing either because of repeated high amplitude sea level falls as postulated by Haq et al. (1988) or because of tectonics. The role of increased paleoproductivity towards the latest Paleocene (Speijer et al., 1996) has to be further evaluated and has to be considered in sea-level reconstructions for this interval.

High plankton values of 90-100% can be found in marls from the early *M. edgari* Zone to *M. subbotinae* Zone (Fig. 7-1). The isolated ‘mfs’ of an incompletely resolved cycle is therefore assumed to be located in the early *M. edgari* Zone. A similar period of high sea level is also described from the Western Desert of Egypt (Luger 1985) and Western Sinai/Negev (Abbass et al., 1994; Speijer, 1994a).

**YpSin-1**

Cycle boundary YpSin-1 is located at the base of partly dolomitic pack- and wackestones which are commonly grouped into the Thebes Formation. The limestones lie directly over marls with high plankton values. Thin section investigations showed that species of the genus *Morozovella* (indicating deeper water conditions, Hart 1980) are completely missing in the limestone so that a sea level drop has to be assumed. The precise age of this sea level fall is hard to determine because of the lack of biostratigraphic data from the limestones. Nevertheless, in the Taba section (M), a 4 m marl interval developed 12 m above the limestone base, was determined as belonging to the *M. formosa formosa* Zone (NP12). The age of the limestone base and cycle boundary therefore is *M. subbotinae* or *M. formosa formosa* Zone (NP11 or early NP12). Because the top of the marls, which is directly below the Thebes limestones, ranges stratigraphically from *M. edgari* Zone to *M. subbotinae* Zone (NP10 to NP11), the limestone might have eroded parts of the underlying marls, creating an unconformity. The diachronous character of the marl - limestone contact has been also described by Flexer (1964) and Kuss (1992).

Lowered sea level during the Thebes Formation is supported by investigations in SW-Sinai (Shahin, 1990). Shahin interprets the limestones of the Thebes Formation in SW-Sinai as inner neritic in contrast to the underlying middle neritic to bathyal marls.

### 7.6.4. Second order supercycles

Reconstruction of 2nd order supercycle boundaries in this study is based on the relative amplitudes assumed for the individual 3rd order sea level drops. The amplitudes serve here as a proxy for sequence stacking patterns which are normally used in supercycle interpretation as introduced by Vail et al. (1977). A 2nd order sequence boundary is assumed for those 3rd order sequence boundaries in Central Sinai which represent the strongest sea level drop in a series of sequences. Because several indicators for the interpretation of sea level fall are used here (mainly P/B ratio, hiatuses and hard beds), a direct comparison of sea level amplitudes in successive sequences is difficult in certain cases. For a better integration of the different data sets, hard beds were interpreted here as being approximately comparable to relatively low planktonic values in the P/B-curve, which for example can be observed in the *Gl. pseudomenardii* Zone.

Three supercycle boundaries were reconstructed for the latest Maastrichtian to early Eocene period (Fig. 7-3). The first 2nd order boundary lies just above the K/T boundary (equivalent to 3rd order sequence boundary DaSin-1). The K/T-hiatus is interpreted as being associated with one of the strongest sea level falls during the studied interval because of its clear regional, erosive character. A second supercycle boundary is interpreted for the *P. pusilla pusilla* / *Gl. pseudomenardii* zonal boundary (equivalent to 3rd order sequence boundary ThSin-2), based on the amplitude -‘stacking pattern’ observed in the P/B ratio and on other sedimentological parameters (see discussion above). Another 2nd order boundary is interpreted to be at the base of the Ypresian limestone package (Thebes Fm.), equivalent to the 3rd order sequence boundary YpSin-1.

### 7.7. Control mechanisms: eustasy vs. tectonics

The geodynamic model of the region suggests that central and southern Sinai were situated in a tectonically rather calm zone during the Mesozoic and Paleogene (see above). The sea level cycles described in this contribution are therefore likely to be attributable to eustatic sea level changes. However, loading and relaxation effects, in terms of a foreland basin model connected to compression in the Syrian Arc, cannot be completely ruled out for central and southern Sinai. In order to test the cycles for eustasy, the cycles were compared to regional sea level curves from Egypt (Luger, 1985; Speijer, 1994b), Tunisia (mainly Kouwenhoven et al., 1997), and Texas (Davidoff & Yancey, 1993), as well as with a ‘global’ sea level curve (Haq et al., 1987) and a Western European sea level curve (Hardenbol et al., in press) (last two columns in Fig. 7-3). Correlation of the different cycles is carried out within the biostratigraphic resolution of planktonic foraminifera and calcareous nannofossils. Moderate discrepancies in the published biostratigraphic ages of characteristic sea-level maxima or minima are normally tolerated in the correlations, as long as the shifts occur within a biozone. In many of these
cases, the seeming differences in age may be caused by different interpretations of biostratigraphic concepts, preservational and paleoecologic problems, or undeciphered hiatuses.

7.7.1. Comparison with Southern and Central Egypt

A comparison with the sea level history as published for Southern and Central Egypt (Luger, 1985; Luger et al., 1997, in press) shows many similarities. Good correlations exist, especially for the latest Maastrichtian (MaSin-Z, DaSin-1), the P. pusilla pusilla - Gl. pseudomenardii zonal boundary (ThSin-2), the late Thanetian (ThSin-5) and the latest Paleocene / early Eocene (period of high sea level) (for details see sequence description above and last column in Table 7-1). Note that Luger (1985) defined the base of his ‘M. velascoensis Zone’ by the first occurrence (FO) of the Acarinina soldadoensis-group. This FO lies within the Gl. pseudomenardii Zone (NP7/8) as defined in the standard Tethys biozonation in Toumarkine & Luterbacher (1985), Blow (1969), and Berggren & Van Couvering (1974), which is used in this study.

A paleodepth curve from the Nile Valley of southern Egypt (Gebel Oweina) was compiled by Speijer (1994b) based on the analysis of benthonic foraminifera. The curve is in good correspondence with the results from Sinai up to at least the early Gl. pseudomenardii Zone. Three of the four sea level falls can be correlated with sequence boundaries interpreted in this contribution. Among these are the sea level drops developed during the Pr. trinidadensis Zone (DaSin-3), M. angulata Zone (ThSin-1) and at the P. pusilla pusilla - Gl. pseudomenardii zonal boundary (ThSin-2). It remains unclear whether the sea level fall interpreted by Speijer (1994b) for the late Gl. pseudomenardii Zone can be correlated with sequence boundary ThSin-5, because Speijer postulates a preceding major sea level high, similar to that indicated in the ‘eustatic’ curve of Haq et al. (1988), for which no evidence can be found in Central East Sinai.

Strougo (1986) described a number of characteristic hiatuses, facies jumps and drastic thickness variations from the Gl. pseudomenardii Zone and M. velascoensis Zone of the Western and Eastern Desert of Egypt. Based on these sedimentary patterns Strougo postulated late Paleocene block faulting associated with the opening of the Gulf of Suez, which by most workers is considered as tectonically inactive until the late Eocene (e.g. Girdler, 1991). At least some of the observations described by Strougo (1986) can be explained by the regional sea level cycles studied in Central East Sinai.

7.7.2. Comparison with El Kef, Tunisia

A comprehensive paleobathymetric study of the El Kef section based on biostratigraphic and paleoecologic investigations of planktonic and benthonic foraminifera, as well as calcareous nannofossils has recently been presented by Kouwenhoven et al. (1997). They reconstructed Paleocene sea level changes for several intervals on the basis of the P/B distribution and verified their findings by quantitative investigations of benthonic foraminifera. In the P/B curve they present from El Kef, several of the characteristic patterns developed in Sinai can be recognized. While the basal Paleocene is reported to be still controlled by relatively high productivity conditions and to possess diagenetically altered horizons, the first major drop in planktonic values can be biostatigraphically correlated with the sequence boundary DaSin-3 in Sinai (in El Kef: within NP3 and Pr. trinidadensis Zone) and is also interpreted as a sea level drop by Kouwenhoven et al. (1997). The next significant P/B drop in the El Kef section is developed around the Pr. uncinitata / M. angulata zonal boundary (within NP4), representing the sequence boundary ThSin-1 in Sinai. This sea level drop with subsequent deepening has been also previously described from Tunisia by Saint-Marc & Berggren (1988) based on quantitative investigations of benthonic foraminifera. The next major drop in planktonic values in the P/B curve from El Kef (Kouwenhoven et al., 1997) is developed in the late P. pusilla pusilla Zone (NP5) and leads to a limestone bed which, similar to our interpretations for Sinai, is considered by Kouwenhoven et al. (1997) as a hardground. A similar limestone with abundant shallow water-indicating (Saint-Marc, 1992) Frondicentaria phosphatica has been described by Salaj (1986) from El Kef. This sea level drop corresponds to the sequence boundary ThSin-2 in Sinai. Within the interval of the Gl. pseudomenardii Zone, the P/B curve of Kouwenhoven et al. (1997) shows two other distinct P/B peaks. The respective drops in planktonic values correlate biostratigraphically well with those in Sinai. Although Kouwenhoven et al. (1997) do not use these P/B changes for paleobathymetric reconstructions, because of possible interference by the increasing paleoproductivity signal, the overall good correspondence in timing may validate a sea level-based interpretation. While the first P/B drop has an NP6 age (ThSin-3), the second drop lies within NP7/8 (ThSin-4). The sequence boundary ThSin-5 and the successive interval cannot be convincingly compared with Sinai, probably due to intensified eutrophication in El Kef or other processes.

The overall good correspondence between many of the Paleocene sea level cycles in El Kef and on Sinai shows that deposition during most of the Paleocene was dominated by basinwide rather than by differential, local processes.
7.7.3. Comparison with Sequences in Eastern Texas

Sea level development of eastern Texas (Davidoff & Yancey, 1993; Mancini & Tew, 1995) is very similar to the sea level history as recorded on Central Sinai (this study) and to the Exxon curve (Fig. 7-6; Haq et al., 1987) at least during the early and middle Paleocene. Texas was situated on a passive continental margin which was gradually subsiding since the middle Mesozoic (Ross & Scotese, 1988; P. V. Heinrich, written communication, 1996). The region was tectonically quiet, except for local movements of salt domes. A maximum flooding surface recorded in the *P. eugubina* Zone of the earliest Paleocene of Texas (Davidoff & Yancey, 1993) may be correlated to the mfs in sequence DaSin-1/DaSin-2 on Sinai. The three sequence boundaries in the *Pr. trinidadensis* Zone, *Pr. angulata* Zone and earliest *Gl. pseudomenardii* Zone of Texas (Davidoff & Yancey, 1993; Mancini & Tew, 1995) can be also found in Sinai (DaSin-3, ThSin-1, ThSin-2). The excellent inter-Tethyan east-west-correlation between Texas and Sinai hints towards a eustatic control on sequence development in both areas during that time.

7.7.4. Comparison with the ‘Global’ and the ‘Western European’ Sea Level History

Many stratigraphic intervals in the Exxon-curve (Haq et al., 1987), especially the Late Cretaceous, are questioned today (e.g. Hancock, 1989, 1993; Miall, 1991, 1992). Furthermore, apparent restrictions to the concept of a single eustatic curve suggest that it is better to consider several regional curves, instead. The data used by Haq et al. (1988) for the Paleogene cycle chart is derived from more than 31 measured sections from at least ten basins on three continents (Armentrout et al., 1993). They provide no citation for any subsurface or seismic analyses used in compiling the cycle pattern.

However, a significant correspondence exists between the global composite chart of Haq et al. (1988) and the major Paleocene sequences of Central Sinai as reconstructed in the present study. Timing, duration and relative amplitude show striking similarities for the cycles DaSin-3/ThSin-1 (long spanning sea level low in *Pr. trinidadensis* Zone / NP3), ThSin-1/2 (sea level drop in *M. angulata* Zone) and the lower part of the composite cycle covering the long sea level low in the *Gl. pseudomenardii* Zone (Fig. 7-6). The sea level drop during the latest Maastrichtian in Central Sinai (MaSin-Z) seems to be also recorded in the Exxon-curve. One main difference concerns the development of the late *Gl. pseudomenardii* Zone, where a sea level low in Central Sinai stands against a sea level high in the Exxon-curve. Another difference between the Sinai and Exxon curve is the three high frequency cycles in Haq et al. from the latest Paleocene (*M. velascoensis* Zone). A comparison between the three second order cycle boundaries reconstructed for Central Sinai and those in the Exxon chart yields varying degrees of correspondence. While the supercycle boundary at the *P. pusilla pusilla* / *Gl.

<table>
<thead>
<tr>
<th>Epochs and Stages</th>
<th>Planktonic Foraminifera biozones</th>
<th>Calcareous nannofossil biozones</th>
<th>Global eustatic curve</th>
<th>Correlation with sequence boundaries reconstructed for Central East Sinai</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Eocene</strong></td>
<td><em>M. foraminiferoides</em></td>
<td>P7</td>
<td>Rising</td>
<td>Good correlation with Haq et al. (1988)</td>
</tr>
<tr>
<td></td>
<td><em>M. subbotinae</em></td>
<td>P6B</td>
<td></td>
<td>Not described by Haq et al. (1988)</td>
</tr>
<tr>
<td></td>
<td><em>M. edgar</em></td>
<td>P6B</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Thanetian</strong></td>
<td><em>P. pseudomenardii</em></td>
<td>P4</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td><em>P. pusilla pusilla</em></td>
<td>P3B</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Paleocene</strong></td>
<td><em>M. angulata</em></td>
<td>P3A</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td><em>P. umbonata</em></td>
<td>P3B</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Danian</strong></td>
<td><em>M. trinidadensis</em></td>
<td>P1C</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td><em>P. sowerbyi</em></td>
<td>P1B</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Maastrichtian</strong></td>
<td><em>A. mayaroensis</em></td>
<td>NC23</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td><em>G. gansseri</em></td>
<td>NC22</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Figure 7-6. Paleocene eustatic sea level curve after Haq et al. (1988) and a correlation to sequence boundaries reconstructed for Central East Sinai in this study. Shaded are 'global' sea level drops which are also recognized in Central East Sinai (sequence boundary numbers as in Fig. 7-3). See text for full discussion.
Geographically, the nearest available, detailed regional sea level study comes from the European Basins (Hardenbol et al., in press). Again, there is good correlation for the sequence boundaries DaSin-3, ThSin-1 and ThSin-2 (sea level drops in Pr. trinidadensis Zone, M. angulata Zone and at the P. pusilla pusilla / Gl. pseudomenardii zonal boundary). Good correspondence of a contemporaneous sea level fall in the Nile Valley of southern Egypt and a regional unconformity in the North Sea Basin during the M. angulata-z (age equivalent to ThSin-1) was already noted by Speijer (1994b; see also the North Sea sequence study of Armentrout et al., 1993). The relatively short sea level drop in the early Paleocene of Sinai (Ps. pseudobulloides Zone) might be equivalent to the Da1-sequence boundary in the European Basins (Hardenbol et al., in press). In addition, the number and timing of the four individual cycles belonging to the long period of low sea level during the Gl. pseudomenardii Zone in Central Sinai correspond well with the sequence boundaries Th1, Th2, Th3 and Th4 shown in the chart of Hardenbol et al. (Fig. 7-3). This is in contrast to the discrepancies occurring in the correlation between the Sinai and the Haq et al. curves indicating that the roles of tectonics and paleoproductivity changes during the late Gl. pseudomenardii Zone and the M. velascoensis Zone have to be further evaluated. It is clear however, that the Exxon-curve does not necessarily represent a global solution for the regional sea level developments.

The high degree of correspondence between the sequences from Sinai and those in Haq et al. (1988) and Hardenbol et al. (in press) provides evidence that hemipelagic deposition for the Paleocene interval of Central Sinai was mainly controlled by eustatic sea level changes at least until the mid-07, pseudomenardii Zone. Furthermore, the relatively similar development of the P/B-curves and the eustatic curve from Haq et al. seems to confirm the usage of the P/B-ratio as a paleobathymetric indicator in this study.

7.8. Conclusions

The middle neritic to upper bathyal Paleocene hemipelagites of Sinai obviously represent a paleobathymetric interval suitable for sea level changes to be recorded by foraminiferal faunal composition. In more pelagic environments similar sea level variations probably cause much smaller to almost no variations in the foraminiferal assemblages and paleoecologically relevant parameters. This is because many physico-chemical parameters change markedly with depth in the upper hemipelagic zone, but show only minor alterations in the deeper hemipelagic and true pelagic zones. In environments shallower than those studied here, benthonics are likely to dominate the foraminiferal fauna. Only extreme sea level highs may be able to import planktonic foraminifera into the depositional area, and, thus, small sea level rises may not be discernible in the P/B-curve. Decreasing water depths may also lead to a more incomplete stratigraphic record and to complications in biostratigraphy because of the absence of biozonal planktonic foraminifera marker species. This study was also able to show that a cyclicity in depth related faunal change is recognizable even in seemingly monotonous deep water facies (see also Brett, 1995: 609). Studies of hemipelagites, therefore, can contribute significantly to deciphering the sequence stratigraphic history and may be considered a valuable tool.

Within the uppermost Maastrichtian to lower Eocene interval in central east Sinai, ten 3rd order sequence boundaries were reconstructed. Sequence stratigraphic history was compared with regional sea level curves from Central- / Southern Egypt, Tunisia and Texas, as well as with the European (Hardenbol et al., in press) and the 'global' (Haq et al., 1987) sequence charts. In general, an excellent correlation is observed in all cases for at least the early and middle Paleocene (K/T to mid Gl. pseudomenardii Zone). Deposition in Central Sinai during this time may therefore have been predominantly controlled by eustatic sea level rather than local tectonics. Sea level development in the late Thanetian is more ambiguous and was possibly influenced by increasing surface water paleoproductivity or tectonics.
8. Comparative biostratigraphy of calcareous nannofossils and planktonic foraminifera in the Paleocene of Eastern Sinai, Egypt

*A. M. Marzouk and §S. Lüning

*Tanta University, Faculty of Science, Geology Department, Tanta 31511, Egypt
§University of Bremen, FB5 - Geosciences, Box 330440, 28334 Bremen, Germany

Abstract

The Paleocene succession in Sinai is characterized by hemipelagic deposits with abundant and well preserved assemblages of planktonic and benthonic foraminifera and calcareous nannofossils. For this contribution, eight sections from central east and north east Sinai were biostratigraphically investigated. Almost all important biostratigraphic index forms and biozones of planktonic foraminifera and calcareous nannofossils have been found so that biozonation within the studied sections, in general, was accomplished without major problems. A comparison of the biozonal distribution of planktonic foraminifera and calcareous nannofossils in the sections yields a consistent regional correlation pattern. However, if compared to the schemes of other authors from other regions, a great variability in relative timing of the foraminiferal and nannofossil bio-events throughout the different studies can be observed. The strong variability of foram-nanno correlations has been described by several authors before. Potential reasons for the shifts observed are discussed in the text. The study suggests that interbasinal and sometimes even intrabasinal correlations using nannofossil and/or foraminiferal biostratigraphic data, must take similar (diachronous) variabilities into account.

8.1. Introduction

Paleocene biozonation schemes for planktonic foraminifera (Blow, 1969; Berggren & Miller, 1988; Berggren et al., 1995) and calcareous nannofossils (Martini, 1971) are well established today. However, their correlation in different regions and even within the same region still yield major discrepancies (e.g. Bolli et al., 1985; Haq et al., 1988; Berggren et al., 1995). In this study, we compare the biozonations of planktonic foraminifera and calcareous nannofossils in eleven Paleocene sections from north east and central east Sinai (Egypt). The Paleocene planktonic foraminifera of Sinai previously attracted attention of several authors because the planktonic foraminifera are relatively abundant (details on planktonic-benthonic foraminiferal ratios in Lüning et al. submitted b) and are mostly well preserved. Triggered by petroleum exploration in the Gulf of Suez, most studies focussed on western Sinai (e.g. Abbass et al., 1994; Abdelmalik et al., 1978a; Anan, 1992; Ayyad & Hamama, 1991; Shahin, 1990) while contributions from northern Sinai (Hewaidy, 1987) and central east Sinai (Said & Kenawy, 1956) are rare. Only a few studies of Paleocene calcareous nannofossils from Sinai have been carried out so far (e.g. Abdelmalik et al., 1978b; Marzouk & Hussein, 1994; Philipsen, 1994; Marzouk & Abou-El-Enein, 1995; Swedan et al., in press). Correlations between biozonation schemes of planktonic foraminifera and calcareous nannofossils from the Paleocene of Sinai are unknown to us.

8.2. Regional Setting

During the Paleocene, Sinai was part of the broad northern shelf margin of the Afroarabian Plate. The Gulf of Suez and Gulf of Aqaba rifts, which bound the Sinai microplate today, were still closed. While Northern Sinai and the adjacent Negev were affected by transpressive movements since the Turonian onwards, Central and Southern Sinai, are thought to have remained tectonically rather calm throughout the whole Mesozoic and Early Tertiary (Said, 1962; Cohen et al., 1990; Kerdany & Cherif, 1990). The SE-vergent, NE-SW trending domal anticlines in Northern Sinai are part of the "Syrian Arc" (Krenkel, 1924, 1925) (Fig. 8-1) which represents an intraplate foldbelt extending from Egypt to Syria formed by Late Cretaceous to recent inversion of Late Triassic / Liassic halfgrabs (Moustafa & Khalil, 1990; Chaimov et al., 1992; Shahar, 1994). Following the nomenclature of Said (1962) the inversion zone can also be called "unstable shelf", and the southern tectonically calm block "stable shelf" (Fig. 8-1). Most of the sections for this study were taken on the stable shelf while three sections are located in synclines on the unstable shelf of northeastern Sinai (sections A8, W) with one of them (section A1) situated directly at the anticlinal flank of the Gebel Areif El Naqa (Fig. 8-1).

At Paleocene times, Central Sinai was part of a northeast-southwest striking intrashelf basin that was confined on the S by the Afroarabian paleoshoreline in the area of the modern Red Sea and on the N by the anticlinal domes.
of the foldbelt (paleogeographic, paleobathymetric and isopach maps in Bartov & Steinitz, 1977: 142; Sestini, 1984: 169; Said, 1990c: 469; Dercourt et al., 1993; Speijer, 1994a: 123). Various studies of foraminifera showed that throughout the Paleocene, a middle to outer shelf environment prevailed in Central Sinai and in the synclinal areas of Northern Sinai (Hewaidy, 1987; Shahin 1990, 1992; Ismail, 1992; Speijer, 1994a: 123; Abbass et al., 1994). The hemipelagic sedimentation was dominated by often bioturbated, greenish chalky foraminiferal marls which are classified as ‘Esna Shale Formation’ (Beadnell, 1905) in Northern and Eastern Sinai (Said, 1990c: 454; Ziko et al., 1993) and as Taqiye Formation in the Negev Desert (Bartov et al., 1972; Bartov & Steinitz, 1977). Similar lithologies in the Paleocene are also developed in the neighbouring countries such as in Tunisia (e.g. Kouwenhoven et al., 1997), Libya (Berggren, 1974) and Syria (Al-Helou, 1996). A model for the Paleocene sea level history of central east Sinai based on microfossil distribution patterns, hiatuses and (rare) lithological changes, has been recently presented by Luning et al. (submitted b).

8.3. Methods

During two field expeditions on Sinai 1995/96, eight sections from the Paleocene marls were measured on the stable shelf and three sections on the unstable shelf with a total of 261 samples (Figs. 8-1 to 8-8). The sections were sampled at intervals ranging between 1-3 metres. To isolate the foraminifera from the matrix, the samples were washed twice with a 63μm sieve after treatment with H2O and the highly concentrated tenside REWOQUAT, respectively. The microfossil residue was then fractionated into four grain size classes for easier handling. The planktonic foraminifera were determined under the light microscope (by Luning). For the calcareous nanoplankton, smear slides were prepared using techniques described in Bramlette & Sullivan (1961) and Hay (1961, 1965). The slides were examined (by Marzouk) under the light microscope at a magnification of about x1250 by both cross-polarized and phase-contrast. The relative abundances of the calcareous nanofossil species were determined as follows: Samples with more than 1 specimen per field of view are listed as ‘abundant’, 1 specimen in up to 10 fields of view as ‘common’, 1 specimen in up to 50 fields of view as ‘few’, and 1 specimen in more than 50 fields as ‘rare’.

8.4. Biostratigraphy

The Paleocene Series consists of three stages (Fig. 8-9): the Danian, Selandian and Thanetian (Jenkins & Luterbacher, 1992, Berggren et al., 1995: 193). While the Danian-Selandian boundary is defined at the P. uncinata-M. angulata boundary, the Selandian-Thanetian boundary lies within the Gt. pseudomauritani Zone and within the nannofossil zone NP6 (Berggren et al., 1995) so that the exact boundary horizon cannot be reconstructed by means of planktonic foraminifera and calcareous nanofossils. In the study area, almost all important biostratigraphic index forms and biozones of planktonic foraminifera and calcareous nanofossils have been found so that biozonation within the studied sections, in general, was accomplished without major problems in terms of preservation or paleoecology. Some minor exceptions are discussed later in the text. Below can be found also some remarks about problems with identification of the different zonal marker species which occur in biostratigraphic practical work. It is clear that more or less precise definitions for all species exist in the literature, nevertheless interpretations of the species definitions vary to some extent from one biostratigrapher to another. The resulting differences may contribute to the discrepancies observed in the correlation of biozonal schemes between planktonic foraminifera and calcareous nanofossils.
A regional unconformity with varying vertical extension is developed around the K/T boundary and is documented by missing biozones and reworked upper Maastrichtian microfauna. Other hiatuses were found within the M. angulata-Igorina pusilla pusilla Zones (sections K, P, R) and the G. pseudomenardii Zone (section T1). The different hiatuses are interpreted to be associated with sea level falls (see discussion in Lüning et al., submitted b).

8.4.1. Planktonic Foraminifera

The biostratigraphic scheme for the planktonic foraminifera of the Paleocene is based on the Tethyan zonal concepts of Blow (1969), Berggren & Van Couvering (1974) and Toumarkine & Luterbacher (1985). The zonation scheme by Berggren et al. (1995) has not been used because our study was started before publication of that scheme. Furthermore, all index species (including e.g. Igorina pusilla pusilla) have been found in the study area so that there was no urgent need to drop the scheme in Toumarkine & Luterbacher (1985). The P-zonation nomenclature is not used in this contribution in order avoid miscorrelations between the different biostratigraphic schemes in the literature which unfortunately use the same P numbers for different-defined biozones (e.g. Blow, 1969; Berggren et al., 1995).

Abathomphalus mayaroensis Zone (Brönnimann, 1952): Interval from the FO of A. mayaroensis to the first occurrence of Paleocene foraminifera. Abathomphalus mayaroensis (Fig. 8-10/6 and 7) is relatively abundant in most parts of this biozone in eastern Sinai, however it is missing in a few horizons, although these horizons contain significant amounts of globotruncanids. In other parts of the region, for example in southern Egypt, A. mayaroensis is completely absent for paleoecologic reasons (e.g. Luger et al. in press). In the study area the A. mayaroensis Zone has been proven in sections A8, A1, C, F and P (Figs. 8-2 to 8-7). Plummerita reicheli [sensu Masters, 1993; synonymous with P. hantkeninoides (Brönnimann) in Robaszynski et al. (1984)] (Fig. 8-10/12 and 13) has been found in the uppermost Maastrichtian of sections A1 and C (Figs. 8-2, 8-4, 8-5). This species is also present in areas where A. mayaroensis is absent, e.g. in southern Egypt (Luger et al. in press; discussion of biozonation herein) and therefore may be helpful for interregional correlations.

Parvularugoglobigerina eugubina Zone (Luterbacher & Premoli Silva, 1964): Interval between FO of P. eugubina and FO of P. pseudobulloides. Parvularugoglobigerina eugubina (Fig. 8-10/1 to 3) has been only found in section C (Fig. 8-5) where the P. eugubina Zone reaches a thickness estimated to be 50-100 cm. Because of the very small dimensions definite identification of P. eugubina needs to be done under the scanning electron microscope. The species has been previously described from the Eastern Desert in Egypt by Luger (1988), Strougo et al. (1992) and Luger et al. (in press). The P. eugubina Zone is reported from western Sinai by Ismail (1992) from the Matulla section, although thickness (approximately 7 m) seems to be overestimated.

Parasubbotina pseudobulloides Zone (Leonov & Alimariana, 1961): Interval from the FO of P. pseudobulloides (Fig. 8-10/4 and 5) to the FO of Praemurica trinidadensis. This biozone is proven in five sections.

Praemurica trinidadensis Zone (Bolli, 1957): Interval from the FO of P. trinidadensis (Fig. 8-10/8 and 9) to the FO of Praemurica uncinata. Because the species P. uncinata develops from P. trinidadensis, all transitions between the two forms exist and, therefore, interpretation of the earliest form of P. uncinata may vary among the different authors. The P. trinidadensis Zone is developed in eight of the eleven sections.

Praemurica uncinata Zone (Bolli, 1957; emended Bolli, 1966): Interval from the FO of P. uncinata (Fig. 8-10/10 and 11) to the FO of Morozovella angulata. Similar to the situation in the biozone below, the interpretation of the earliest form of M. angulata may vary from author to author because the chambers of M. angulata evolve in a continuous manner from rounded to angular and identification of the oldest M. angulata depends on the (to some extend subjective) critical angularity value used. The P. uncinata Zone is developed in all sections.

Morozovella angulata Zone (Bolli, 1963): Interval from the FO of M. angulata (Fig. 8-10/14 and 15) to the FO of Ig. pusilla pusilla. This zone has been found in seven of the eleven sections.

Igorina pusilla pusilla Zone (Bolli, 1957): Interval from the FO of Ig. pusilla pusilla (Fig. 8-11/1 and 2) to the FO of G. pseudomenardii. The upper zonal boundary may be alternatively interpreted using the FO of M. velascoensis which is virtually contemporaneous to the FO of Gl. pseudomenardii. A strong regional sea level drop around the Ig. pusilla pusilla / Gl. pseudomenardii zonal boundary (see Lüning et al. submitted b) locally lead to a significant decrease in abundance of the marker species Gl. pseudomenardii and M. velascoensis which complicates reconstruction of this biozonal boundary. While in the study area the two marker species can be still found just above the biozonal boundary, in shallower areas, for example in the Galala Mountains at the
Figs. 8-2 to 8-4. Correlation of studied Paleocene sections by planktonic foraminiferal biozones. The projected transect starts in the NW (left of Fig. 8-2) and is continued towards the SE in Figs. 8-3 and 8-4 (location map in Fig. 8-1). Legend in Fig. 8-4.
Section C near Sheikh Attiya

Legend

Columns in sections:

<table>
<thead>
<tr>
<th>Plankton foram biozones</th>
<th>Calcareous nannofoss. biozones</th>
<th>Lithologic column</th>
</tr>
</thead>
</table>

Lithologic symbols:

- limestone
- calcareous marl
- marl
- shale
- dolomite
- shaley calcareous marl
- chalk
- shaley chalk

Biozone boundaries:

- exact biozone boundary
- approximate boundary or biozone data missing beyond this boundary
- hiatus / unconformity
- first occurrence of Plummerita reicheli
- relative sequence of biozonal boundaries 1-2 and X-Y interpreted from relationships observed in other sections (see further explanations in Fig. 8-14)

Fig. 8-4
Figs. 8-5 to 8-8. Planktonic-benthonic ratio, range charts for planktonic foraminifera (non-quantitative) and distribution charts for calcareous nannofossils (semiquantitative). Legend for relative abundances of calcareous nannofossils in Fig. 8-8; legend for lithologies in Fig. 8-4. 8-5 Section C, 8-6 Section F, 8-7 Section A8, 8-8 Section R.
<table>
<thead>
<tr>
<th>Planct.</th>
<th>Calcarr.</th>
<th>benth.</th>
</tr>
</thead>
<tbody>
<tr>
<td>foramin. zones</td>
<td>nanofossils zones</td>
<td>100%</td>
</tr>
</tbody>
</table>

Fig. 8-7: Section A8
3o
O
Paleocene
Selandian Thanetian
Eocene
Ypresian
Esna Shale

Parasubb pseudobulloides
Praemurica trinidadensis
Praemurica uncinata
Morozovella angulata
Morozovella conicotruncata
Planorotalitas chapmani
Igorina pusilla pusilla
Morozovella acuta
Morozovella velascoensis
Globanom pseudomenardii
Morozovella edgari
Coccolithus pelagicus
Thoracosphaera operculata
Placozygus sigmoides
Cruciplacolithus primus
Cruciplacolithus tenuis
Chiasmolithus danicus
Ericsonia subpertusa
Ellipsolithus macellus
Chiasmolithus consuetus
Fasciculithus bitectus
Fasciculithus tympaniformis
Fasciculithus ulii
Fasciculithus janii
Neochiastozygus perfectus
Fasciculithus billii
Heliolithus cantabriae
Heliolithus kleinpellii
Neococcolithes protenus
Discoaster bramletti
Discoaster mohlen
Discoaster okadai
Discoaster muttiradiatus
Fasciculithus involutus
Fasciculithus alanii
Discoaster nobilis
Discoaster falcatus
Discoaster lenticulatus
Thbrachiatus nunnii
Discoaster elegans
Discoaster diastypus
Rhomboaster bitnlida
Ericsonia tormosa
Discoaster barbadiensis
Discoaster binodosus
Tribrachiatus orthostylus
Ellipsolithus distichus
Pontosphaera multipora
Chiasmolithus cahfonicus
Chiasmolithus eograndis
Neochiastozygus junctus

Fig. 8.8: Section R

Paleocene
Selandian Thanetian
Esna Shale

Legend for
carbonate microfossils
Fig. 8-9. Correlation of biozonal schemes of planktonic foraminifera and calcareous nannofossils in Eastern Sinai and comparison with correlations from the literature. Note that for graphical reasons, the nannofossil zones were correlated with a ‘fixed’ foraminiferal zonal scheme. Absolute ages taken from Berggren et al. (1995) and correlated with planktonic foraminiferal biozones.
Fig. 8-10. Biostratigraphically important planktonic foraminifera from the latest Maastrichtian and Paleocene of eastern Sinai. Scale bars = 100 μm.

1-3: Parvularugoglobigerina eugubina (Luterbacher & Premoli Silva), section C, all specimens from sample C1-93.
4-5: Parasubbotinita pseudobulloides (Plummer), section F, both specimen from sample F1-23.
6-7: Abathomphalus mayaroensis (Bolli), section A7, specimen 6 from sample A7-11, specimen 7 from sample A7-15.
8-9: Praemurica trinidadensis (Bolli), section F, both specimens from sample F1-30.
10-11: Praemurica uncinata (Bolli), section F, both specimens from sample F1-39.
12-13: Plummerita reicheli (Brönnimann) [sensu Masters, 1993], section A5, both specimens from section A5 (near sections A1 and A8).
14-15: Morozovella angulata (White), section F, both specimens from sample F1-42.

Fig. 8-11. Biostratigraphically important planktonic foraminifera from the Paleocene of eastern Sinai. Scale bars = 100 μm.

1-2: Igorina pusilla pusilla (Bolli), specimen 1 from section F, sample F1-42; specimen 2 from section Q, sample Q1-12.
3-4: Globanomalina pseudomenardii (Bolli), section F, both specimens from sample F1-45.
5-6: Morozovella velascoensis (Cushman), specimen 5 from section F, sample F1-46; specimen 6 from section C, sample C1-33.
7-8: Acarinina soldadoensis soldadoensis (Bronnimann), section K, both specimens from sample K1-19.

Fig. 8-12. Calcareous nannofossils from the Paleocene of eastern Sinai. Scale bar = 15 μm.

4: Chiasmolithus consuetus (Bramlette & Sullivan) Hay & Mohler, section F, Gebel Misheiti, sample F1-43.
6-8: Cruciplacolithus tenis (Stradner) Hay & Mohler, section F, Gebel Misheiti, sample F1-26.
9, 13, 14: Fasciculithus tympaniformis Hay & Mohler, section F, Gebel Misheiti, sample F1-44.
12: Fasciculithus hitecctus Romein, section F, Gebel Misheiti, sample F1-41.

Fig. 8-13. Calcareous nannofossils from the Paleocene of eastern Sinai. Scale bar = 15 μm.

4: Tribrachiatus orthostylus Shamrai, section C, Sheikh Attiya, sample C1-95.
7, 15: Tribrachiatus contortus (Stradner) Bukry, section F, sample F1-51.
8: Discocysta barbadiensis Tan Sin Hok, section A8, Gebel Areif El Naqa, sample A8-23.
10, 12: Discocysta bipinata Martini, section P, Wadi Gureis, sample P1-35.
14: Rhombaster spinicos (Shafik & Stradner) Perch-Nielsen, section C, Sheikh Attiya, sample C1-35.
northwestern surroundings of the Gulf of Suez (Kulbrok, 1996), *G. pseudomenardii* and *M. velascoensis* are absent so that the upper boundary of the *Ig. pusilla pusilla* Zone cannot be reconstructed properly. The *Ig. pusilla pusilla* Zone has been reconstructed in eight sections.

**Globanomalina pseudomenardii** Zone (Bolli, 1957): Interval between the FO and LO of *G. pseudomenardii* (Fig. 8-11/3 and 4). Because the *G. pseudomenardii* Zone is characterized by a phase of relatively low sea level (Lüning et al. submitted b), planktonic foraminifera including the index species *G. pseudomenardii* become rare or are even absent in some horizons of this zone which poses problems for the reconstruction of the upper boundary of the *G. pseudomenardii* Zone. While for sections K, M, Q, and R (Figs. 8-3, 8-4) it was possible to determine the *G. pseudomenardii-M. velascoensis* zonal boundary, this could not be accomplished for sections F and W (Fig. 8-2) due to preservational-paleoecological reasons. Similar problems are reported from the Eastern Desert where Luger (1985) used the FO of the *Acarinina soldadoensis*-group (Fig. 8-11/7 and 8) to define the upper limit of the *G. pseudomenardii* Zone. The *G. pseudomenardii* Zone has been found in ten sections and can be further subdivided by calcareous nannofossil zones (see below).

**Morozovella velascoensis** Zone (Bolli, 1957): Interval from the LO of *G. pseudomenardii* and the LO of *M. velascoensis*. Because of the relatively high sea level around the upper boundary of this zone with high numbers of planktonic foraminifera, the LO of *M. velascoensis* (Fig. 8-11/5 and 6) can be determined with some confidence from the preservational and paleoecological point of view. The thickness of the *M. velascoensis* Zone seems to be anomalously reduced. A hiatus is recognized in section T1 (Fig. 8-4) and probably in C (Figs. 8-4, 8-5) and P (Fig. 8-3) so that the sedimentary record in this zone may be incomplete throughout the study area. The P/E boundary in this contribution is placed at the boundary between the *M. velascoensis* and *M. edgari* Zones. Because of some discrepancies involved in the inter-continental reconstruction of the P/E boundary it is currently discussed to redefine this boundary and place it at a strong negative change in δ13C which has been observed during the latest *M. velascoensis* Zone at many places in the world (e.g. Kennett & Stott, 1991; Stott et al., 1996).

### 8.4.2. Calcareous Nannofossils

The biostratigraphic scheme for the calcareous nannofossils is based on the zonal concept of Martini (1971).

**Micula prinsii** Zone (Perch-Nielsen, 1979): Interval from FO of *M. prinsii* (Fig. 8-12/1) to FO of *Markhalius inversus*. This zone is developed in sections A8, C, F, P (Figs. 8-2 to 8-6).

**Markhalius inversus** Zone (NP1) (Mohler & Hay in Hay et al., 1967; emended Martini, 1970): Interval from FO of *M. inversus* to FO of *Cruciplacolithus tenuis*. NP1 has been proven in sections C, F and P (Figs. 8-2 to 8-6).

**Cruciplacolithus tenuis** Zone (NP2) (Mohler & Hay in Hay et al., 1967; emended Martini, 1970): Interval from FO of *C. tenuis* (Fig. 8-12/6 to 8) to FO of *Chiasmolithus danicus*. The biozone is developed in five sections.

**Chiasmolithus danicus** Zone (NP3) (Martini, 1970): Interval from FO of *C. danicus* to FO of *Ellipsolithus macellus*. It is interesting to note in this respect that *C. danicus* (Fig. 8-12/2 and 3) shows moderate to strong fluctuations in abundance in most of the studied sections (e.g. section F, Fig. 8-6) which may indicate a certain susceptibility of this species towards paleoecologic changes. This may also lead to regional differences in the FO of *C. danicus* and therefore in the reconstruction of the base of NP3.

**Ellipsolithus macellus** Zone (NP4) (Martini, 1970): Interval from the FO of *E. macellus* to FO of *Fasciculithus tympaniformis*. The boundary between NP3 and NP4 is generally regarded as chronostratigraphically unreliable owing to the assumed susceptibility of the coccolith *E. macellus* (Fig. 8-12/21 to 23) towards dissolution or to environmental restriction of the species (e.g. Monechi & Thierstein, 1985; Wei & Wise, 1989). A detailed discussion on this matter is included in Berggren et al. (1995). NP4 is found in nine sections.

**Fasciculithus tympaniformis** Zone (NP5) (Mohler & Hay in Hay et al., 1967): Interval from FO of *F. tympaniformis* (Fig. 8-12/9, 8-12/13 and 14) to FO of *Heliolithus kleinpellii*. The biozone NP5 has been found in seven sections. Concerning the upper limit of NP5, Wei & Wise (1989) described the FO of *H. kleinpellii* to be diachronous while Berggren et al. (1995) suggested that it is non-diachronous.

**Heliolithus kleinpellii** Zone (NP6) (Mohler & Hay in Hay et al., 1967): Interval from FO of *H. kleinpellii* (Fig. 8-12/19 and 20) to FO of *Discoaster mohleri*. This biozone has been found in seven sections.
Discoaster mohleri Zone (NP7/8) (Romein, 1979): Interval from the FO of *D. mohleri* to FO of *Discoaster multiradiatus*. This zone represents a combined zone including NP7 and NP8 because the index form for NP8, *Heliolithus riedelii*, is absent in the study area. The *D. mohleri* Zone was proven in nine sections.

Discoaster multiradiatus Zone (NP9) (Bramlette & Sullivan, 1961; emended Martini, 1971, and Bukry & Bramlette, 1970): Interval from FO of *D. multiradiatus* (Fig. 8-13/11) to FO of *Tribrachiatus nunnii* and/or *Discoaster diastypus*. NP9 has been found in eight sections.

*Tribrachiatus contortus* Zone (NP10) (Martini, 1971): Interval from FO of *Tribrachiatus nunnii* (Fig. 8-13/3) or *Discoaster diastypus* (Fig. 8-13/2) to LO of *T. contortus* (Fig. 8-13/7, 8-13/15). This biozone has been proven in ten of the eleven sections.

### 8.4.3. Biozonal correlation of planktonic foraminifera and calcareous nannofossils

A comparison of the biozone distribution of planktonic foraminifera and calcareous nannofossils in the studied sections yields a consistent correlation pattern (Fig. 8-9). Sample resolution in most cases is high enough to reconstruct the relative sequence of foraminifera and nannofossil biozone boundaries properly. However, occasionally both the foraminiferal and the nannofossil zonal boundaries lie between the same pair of samples so that the relative succession of the two boundaries is not resolved and therefore remains ambiguous (Fig. 8-14). In these cases, the sequence of bio-events has been interpreted based on clearly developed relationships observed in other sections. It has to be mentioned that no contradictions in terms of relative boundary timing have been observed throughout the studied sections.

Fig. 8-14. If both the foraminiferal and the nannofossil boundaries lie between the same pair of samples, the relative succession of the two boundaries is not resolved and therefore remains ambiguous (scenario 1 and 2). In these cases, the sequence of bio-events has been interpreted based on clearly developed relationships observed in other sections.

The latest Maastrichtian *M. prinsii* Zone correlates with the uppermost *A. mayaroensis* Zone. The absence of *M. prinsii* and the presence of *P. reicheli* in section A1 (Fig. 8-2) may indicate that the FO of *P. reicheli* occurs slightly before the FO of *M. prinsii*. However, *P. reicheli* is missing in other sections where *M. prinsii* in turn is present (sections A8, F, P) so that distribution of at least one of the two species is closely related to paleoecological or preservational aspects. The detailed correlation of the NP zones (Martini, 1971) with the foraminiferal biozones for the study area is included in Fig. 8-9. The comparison of our scheme with those of other authors (Haq & Aubry, 1981; Bolli et al., 1985; Haq et al., 1988; Toker, 1989; Berggren et al., 1995) shows a great variability in relative timing of the foraminiferal and nannofossil bio-events throughout the different studies (Fig. 8-9). For graphical reasons, the nannofossil zones were correlated with a 'fixed' foraminiferal zonal succession, although it is clear that also the foraminiferal biozones may be in part diachronous and therefore contribute to the observed discrepancies. While our NP1/NP2 boundary seems to correspond to the scheme for northern Africa and the Middle East by Haq & Aubry (1981) and lies only slightly higher than that in Haq et al. (1988) and Berggren et al. (1995), the NP2/NP3 boundary in Sinai which lies within the *P. trinidadensis* Zone is significantly younger than in the other studies. Our NP3/NP4 boundary for Sinai matches well the situation described for Turkey (Toker, 1989) while the global schemes of Bolli et al. (1985).
Haq et al. (1988) and Berggren et al. (1995) suggest an earlier and the scheme of Haq & Aubry (1981) a later date for this bio-event. From the NP4/NP5 boundary upwards our zonal correlation corresponds quite well with that in Berggren et al. (1995). The same applies to the schemes in Haq & Aubry (1981) and Haq et al. (1988) with the exception of the NP10/NP11 boundary which we for Sinai and Berggren et al. (1995) for the Tethys assume to be younger. The P/E boundary, as defined here by foraminifera at the M. velascoensis-M. edgari zonal boundary, in our scheme lies within NP10 which corresponds to Berggren et al. (1995) while Bolli (1985) interpreted the P/E boundary to be associated with the NP9/NP10 boundary. In the Haq et al. (1988) correlation the P/E boundary lies within NP9.

The strong variability of foram-nanno correlations has been observed by many authors before (e.g. Monechi & Thierstein, 1985; Varol, 1989; Berggren et al., 1995). Discrepancies in magnetobiostratigraphic correlations with calcareous nannofossils have been discussed in some detail by Berggren et al. (1995: 174ff). They suggest that the discrepancies in correlation are mainly caused by undeciphered unconformities in the stratigraphic record and to a lesser extent by differences in taxonomic concepts of certain calcareous nannofossils. Nevertheless, they also report that diachrony of several million years may occur in certain cases. The statistical reliability of biostratigraphic events is discussed in Monechi & Thierstein (1985). Because of the problems seemingly associated with the scheme of Martini (1971), an alternative biozonation scheme of calcareous nannofossils for the Paleocene had been presented by Varol (1989). This scheme is based on last occurrence dates which are important for studies based on cutting samples which are common in the oil industry.

On the basis of observations about the role of taxonomy, paleoecology and preservation (as listed above in the descriptions of the different biozones) we conclude that the discrepancies in foram-nanno correlations are related to a complex set of parameters. Potential parameters include paleoecological conditions (e.g. water depth, surface productivity), microfossil dissolution and differences in taxonomic concepts (see above). Also intensity of search for index forms may occasionally play a role when index forms are rare. Undeciphered hiatuses may cause changes in the position of a nannofossil bio-event within a foraminiferal biozone, however, they cannot lead to inversion of a pair of biozonal boundaries. Quantitative differentiation of the different parameters for the different biozonal pairs is complicated and would need a higher resolution, stratigraphically more focussed fully (semi-) quantitative study, probably involving sections from several regions. It may be also helpful to complement the biostratigraphic datasets from planktonic foraminifera and calcareous nannofossils with magnetostratigraphic polarity data which would allow a more comprehensive comparison of the different schemes.

8.5. Conclusions

The Paleocene succession in Sinai is characterized by hemipelagic deposits with abundant and well preserved assemblages of planktonic and benthonic foraminifera and calcareous nannofossils. For this contribution, eight sections from central east and north east Sinai were biostratigraphically investigated. Almost all important biostratigraphic index forms and biozones of planktonic foraminifera and calcareous nannofossils have been found so that biozonation within the studied sections, in general, was accomplished without major problems. A comparison of the biozonal distribution of planktonic foraminifera and calcareous nannofossils in the sections yields a consistent regional correlation pattern. However, if compared to the schemes of other authors from other regions, a great variability in relative timing of the foraminiferal and nannofossil bio-events throughout the different studies can be observed. The strong variability of foramin-nanno correlations has been described by several authors before. Potential reasons for the shifts observed are discussed in the text. The study suggests that interbasinal and sometimes even intrabasinal correlations using nannofossil and/or foraminiferal biostratigraphic data, must take similar (diachronous) variabilities into account.
9. Sedimentary response to basin inversion: Mid Cretaceous - early Tertiary pre- to syndeformational deposition at the Areif El Naqa anticline (Northern Sinai, Egypt)

S. Lüning • J. Kuss • M. Bachmann • A. M. Marzouk • A. M. Morsi

S. Lüning (✉) • J. Kuss • M. Bachmann
University of Bremen, FB5 - Geosciences, Box 330440, 28334 Bremen, Germany

A. M. Marzouk
Tanta University, Faculty of Science, Geology Department, Tanta 31511, Egypt

M. Morsi
Geology Department, Faculty of Science, Ain Shams University, Cairo, Egypt

Abstract

The Areif El Naqa domal anticline in northeastern Sinai is part of the ‘Syrian Arc’ which represents an intraplate orogen that has been formed since the late Cretaceous by inversion of an older halfgablen system as a consequence of the collision of the African and Eurasian Plates. The pre- and syndeformational upper Albian to lower Eocene sedimentary succession at the anticline was formed predominantly under calcareous shallow marine to hemipelagic conditions with subordinate siliciclastic intercalations. The depositional history at Areif El Naqa has been reconstructed in terms of sequence stratigraphy on the basis of detailed sedimentologic, biostratigraphic and paleoecologic investigations of ten sections as well as literature data. Following a late Triassic-early Cretaceous extensional period, the Albian-Turonian successions were deposited under technically rather quiet conditions. Inversion started around the earliest Senonian. Three significant uplift phases during the studied period have been determined for the Areif El Naqa anticline based on evidence from lateral facies and thickness changes, local development of pronounced hiatuses and comparisons with the sequence stratigraphic development in the tectonically quiet region of central east Sinai. The first major compressional phase is interpreted to have taken place during the Coniacian-early Santonian and is characterized by pronounced facies and thickness changes which were documented in an earlier study by Bartov et al. (1980). Nevertheless, inter-regional sea level changes still played a significant role for deposition at Areif El Naqa during this period. The second tectonic phase is indicated for the late Campanian-early Maastrichtian by siliciclastics which are interpreted to have been reworked from older siliciclastics uplifted in the anticlinal core. The third compressional period is assumed for the middle Paleocene to early Eocene as evidenced by a significant hiatus in sections at the northern anticlinal flank. The uplift history at Areif El Naqa has been compared with the tectonic development in other parts of the Syrian Arc and in general seems to reflect major movements which occurred throughout the anticlines of the foldbelt.

9.1. Introduction

The Syrian Arc is considered as an intraplate orogen that has been formed since the late Cretaceous by inversion of a late Triassic-early Cretaceous halfgablen system. The Areif El Naqa represents one of the highest domal anticlinal structures in the southern part of the anticlinal system with the exposure of a thick stratigraphic succession of Triassic to middle Eocene carbonates and siliciclastics (Fig. 9-1A to C). Reconstruction of the detailed uplift history of this structure contributes to a better understanding of the deformational development of the Syrian Arc foldbelt and subsequently of the collisional history between the African and Eurasian Plates. This is especially important for the evaluation of controlling mechanisms on sedimentation, mainly for the differentiation between eustasy and regional tectonics. A comprehensive study of the Areif El Naqa has been previously carried out by Bartov et al. (1980) and was mainly based on mapping, lithostratigraphy and macrofossil biostratigraphy. A lithostratigraphic summary for the anticline has been published by Allam & Khalil (1988). The detailed models of the depositional and deformational history presented by Bartov et al. (1980) show that syntectonic sedimentation in the late Cretaceous and early Tertiary was characterized by pronounced lateral facies and thickness differentiation with sedimentary units in general thinning out towards the anticlinal core.
Fig. 9-1
In this contribution we focus on the mid-Cretaceous to lower Eocene succession of the Areif El Naqa and present a biostratigraphic frame based on benthonic foraminifera, calcareous algae, ostracods and ammonoids for the upper Albian-Cenomanian successions, and on planktonic foraminifera, calcareous nannofossils and ostracods for the Turonian-lower Eocene units. The depositional history is reconstructed in terms of sequence stratigraphy and is based on detailed facies and microfacies studies. In order to differentiate between regional (partly eustatic) sea level movements and local shallowing trends caused by tectonic uplift, we compare our results for the anticline with detailed regional sequence stratigraphic studies from the Albian-Cenomanian of northern Sinai (Bachmann & Kuss, submitted), the Turonian-Maastrichtian of central east Sinai (Lüning et al., submitted a) and the Palaeocene of central east Sinai (Lüning et al. submitted b). It has to be mentioned that the Cretaceous-Paleogene deposits in northern Sinai play a significant role in hydrocarbon plays of the (rare) oil and gas discoveries in this region (see Cohen et al., 1990, Alsharhan & Salah, 1996).

9.2. Geodynamic setting

According to Said (1962) the Sinai Peninsula can be subdivided into the tectonically ‘unstable shelf’ of northern Sinai which is dominated by domal anticlines of the Syrian Arc Foldbelt and the ‘stable shelf’ of central and southern Sinai which lacks these features and is characterized by mainly flat lying strata (Fig. 9-1B). The Syrian Arc extends from Northern Egypt to Syria and consists of the Sinai-Negev Fold Belt in northern Egypt and Israel and the Palmyride Fold Belt in Syria (Fig. 9-1A). The orogen represents an intra-continental foldbelt which has been formed by inversion of halfgrabens from the late Turonian / Coniacian onwards (Moustafa & Khalil, 1990). The sediments studied at Areif El Naqa were deposited during the pre- and syninversional phases. The system of ENE striking halfgrabens established when the Turkish-Apulian terrane rifted off the Northeast African plate margin during the Late Triassic / Liassic and drifted towards the north. The southernmost Mesozoic halfgraben is interpreted to lie at the E-W striking Themed Fault (Fig. 9-1B) which marks the boundary between the unstable and stable shelves (Moustafa & Khalil, 1994). The Syrian Arc compression is characterized by dextral transpression and is related to the collision of the Arabian and Turkish Plates in the area of southern Turkey and neighbouring areas along the Bitlis-Zagros Suture (Fig. 9-1A). The compressive stress has been transferred within the halfgraben systems over several hundreds of kilometers to southern directions. Because of structural and lithological similarities it can be assumed that folding of the Sinai-Negev Foldbelt and the Palmyride Foldbelt in general took place contemporaneously and in a similar style (Chaimov et al., 1992; Shahar, 1994). The Sinai-Negev- and Palmyride Foldbelts, therefore, are interpreted to form a single tectonic unit. Discrepancies in the deformational history of both foldbelts and of different anticlines may be explained, at least in part, with their different distances to the Afroarabian-Eurasian collisional zone. Furthermore, the complex geometries of the former graben system may have lead to different styles of deformation, including transpression, transtension, pull-apart-structures and pure strike-slip. Hirsch et al. (1995b) assume that the zone of maximum deformational intensity during the deformational history has steadily migrated northward, from the northern Egyptian craton towards the active collisional zone. Summaries of the tectono-sedimentary basin development of the Eastern Mediterranean (Levantine Basin) were e.g. published by Sestini (1984), Abu-Jaber et al. (1989), Cohen et al. (1990) and Robertson et al. (1996).

9.3. Materials and Methods

Within the course of four field expeditions between 1994 and 1996, ten sections from the Albian to lower Eocene were measured at the Gebel Areif El Naqa anticline in northeast Sinai (Fig. 9-1C). Another 26 sections from the same stratigraphic intervals were studied at other places in northern Sinai (Bachmann et al., 1996; Bachmann & Kuss, submitted) and in central east Sinai (Lüning et al., submitted a, b) which are used here as reference sections from outside the Areif El Naqa area (see below). During fieldwork, special attention was paid to sedimentary structures, facies and horizons with abrupt lithological changes. For thin section analysis, 103 handspecimens from limestones and sandstones were chosen. While a semiquantitative microfacies analysis was carried out for the Albian-Cenomanian (by Bachmann), qualitative microfacies data was collected for the Turonian-Maastrichtian (by Lüning). For microfossil investigations, chalks, marls and shales were sampled at intervals...
ranging between 1-3 metres with a total of 181 samples. For extraction of foraminifera and ostracods the pelitic samples were washed twice using a 63μm sieve after treatment with H₂O₂ and the highly concentrated tenside REWOQUAT, respectively. The washed microfossil residue was fractionated into four grain size classes for easier handling. The planktonic foraminifera were identified under the light microscope (by Lüning). For the calcareous nannoplankton, smear slides were prepared using techniques described in Bramlette & Sullivan (1961) and Hay (1961, 1965). The slides were examined (by Marzouk) under the light microscope at a magnification of about x1250 by both cross-polarized and phase-contrast. Ostracodes were investigated by Morsi and Bassiouni and calcareous algae by Kuss.

9.4. Biostratigraphy

Biostratigraphy in this study is based on planktonic and benthonic foraminifera, calcareous nannofossils, ostracods and calcareous algae, while macrofossils yielded some supplementary data. In the mostly shallow marine to restricted environments of the late Albian to Coniacian where planktonic foraminifera and calcareous nannofossils are rare, biostratigraphy is based on benthonic foraminifera, calcareous algae, ostracods, and ammonites. The biostratigraphic value of the Albian-Cenomanian benthonic foraminifera and calcareous algae has been previously demonstrated by Schroeder & Neumann (1985) while Kuss & Conrad (1991) and Kuss (1994) presented stratigraphic ranges of calcareous algae in the region. Detailed ostracod biozonation schemes have been proposed for the Cenomanian-Turonian (Rosenfeld and Raab 1974) and Coniacian-Maastrichtian (Honigstein, 1984) in Israel. Because some important marker species have not been recorded at Areif El Naqa, biozonation by ostracods in the studied sections is of lower resolution than in the schemes from Israel, however, occasionally provides valuable information. During fieldwork, few ammonites have been found in the late Albian (Geyer et al. in press) and Turonian parts of the succession. However, other studies in which a focussed search for ammonites in the field had been carried out, showed that ammonites may provide a valuable biostratigraphic tool in the Upper Cretaceous of Sinai (e.g. Lewy, 1975; Lewy & Raab, 1976; Abdel-Gawad, 1990; Abdel-Gawad et al., 1996, 1997) and especially in the Gebel Areif El Naqa area (Bartov et al., 1980). The presence of slight discrepancies in the correlation of ammonite and microfossil biozones with the stage boundaries have to be taken into account when using biostratigraphic data from both fossil groups.

Due to paleoecologic reasons, biostratigraphy on the basis of planktonic foraminifera and calcareous nannofossils is restricted to the more hemipelagic parts of the succession. The facies of the upper Coniacian to Santonian and upper Maastrichtian to lower Eocene are often well suited, whereas intense silicification restricts microfossil preservation in the mostly hemipelagic Campanian units. Planktonic foraminifera of the Upper Cretaceous to Lower Tertiary sedimentary succession of Sinai have previously been biostratigraphically studied by several authors. Especially the late Maastrichtian-early Eocene interval attracted attention because here the planktonic foraminifera are relatively abundant and are mostly well preserved. Triggered by petroleum exploration in the Gulf of Suez, most studies focussed on western Sinai (e.g. Abdelmalik et al., 1978a; Cherif et al., 1989a, b; Ayyad & Hamama, 1991; Anan, 1992; Shahin, 1992; Ismail, 1993; Abbass et al., 1994) while contributions from central east Sinai (Said & Kenawy, 1956; Shahin & Kora, 1991; Orabi, 1992, 1993; El Sheikh, 1995; Lüning et al., submitted b; Marzouk & Lüning, in press) and northern Sinai (Hewaidy, 1987; Hewaidy et al., 1991; Hewaidy & El Ashwah, 1993; Sweidan et al., in press) are relatively rare. Only recently, a biostratigraphic study based on Cretaceous planktonic foraminifera has been published for Gebel Areif El Naqa (Ayyad et al., 1996). Unfortunately, they failed to report the exact locations for the different parts of their composite section (Ayyad et al., 1996: Fig. 1) which would be extremely important in the laterally highly variable syntectonic sedimentary system of the Areif El Naqa anticline. The northern one of the two measured sections of Ayyad et al. (1996) is located close to our sections A1a, A1b, A9 (Fig. 9-1C). The comparison with our findings yields several discrepancies. Ayyad et al. (1996: Fig. 3, p.267) describe a 300m thick ‘Cenomanian’ unit which according to previous studies (Bartov et al., 1980) in the region and the presence of Knocmiceras ubigii subcompressum in the lower part of section A10 (Fig. 9-6) (see discussion below) should better be subdivided into a lower part of late Albian age (spanning units M1 and M2 of Ayyad et al., 1996: Fig. 3) and an upper part of Cenomanian age (spanning units M3 - M4). Furthermore, none of the planktonic foraminifera described by Ayyad et al (1996) have been found in our washed samples which matches well the rather shallow paleoenvironment reconstructed for the Areif El Naqa area during this time (see facies distribution in Figs. 9-6, 9-9). In a similar way, also for the Santonian-Maastrichtian the stated location for their northern section is doubtful because exposure at that place is incomplete, their thickness data often do not match our observations by far, and they report the presence of biozones which are unlikely to occur at least at the stated location, even if one considers the strong potential lateral facies changes. Therefore, the contribution of Ayyad et al. (1996) seems to be only of limited use for our study and their data are only referred to in a few selected cases. In our contribution we use the Tethyan planktonic foraminiferal biozonation schemes of Caron (1985) for the Turonian to early Coniacian, of Robaszynski et al. (1984) for the late Coniacian to Maastrichtian and of Toumarkine and Luterbacher (1985) for the Paleocene to early Eocene.

Only a few studies of Cretaceous to Paleogene calcareous nannofossils from Sinai have been carried out so far (Abdelmalik et al., 1978b; Araf & El Ashwah, 1988; Araf, 1991; Philipsen, 1994; Marzouk & Hussein, 1994;
Marzouk & Abou-El-Enein, 1995; El Sheikh, 1995). Biozonation of calcareous nannofossils in this contribution is based on concepts of Crux (1982) for the Turonian to early Campanian, of Eshet & Moshkovitz (1995) and Perch-Nielsen (1979) for the late Campanian to Maastrichtian and of Martini (1971) for the Paleocene-early Eocene. Discussions of the biostratigraphic schemes for calcareous nannofossils and planktonic foraminifera of the Late Cretaceous and Paleocene in Eastern Sinai are included in Lüning et al. (submitted b) and Marzouk & Lüning (in press).

9.5. Facies analysis and paleobathymetry

9.5.1. Shallow marine and terrestrial deposits

The facies interpretations in shallow marine and terrestrial environments are based on the presence of characteristic lithologies (e.g. sandstones, bedded gypsum, carbonates), sedimentary structures (e.g. cross-beded carbonates and sandstones) and characteristic biotic and abiotic components (e.g. rudists, oolites, laminated algal mats, green algae, monospecific hypersaline ostracod faunas). For the Albian-Cenomanian of northern Sinai facies models have been presented by Kuss & Schlagintweit (1988) and Kuss (1992b). New detailed studies of the Aptian-Cenomanian units allow to interpret sea-level fluctuations from facies changes (Bachmann et al., 1996). Furthermore, a sequence stratigraphic interpretation of the late Aptian-Cenomanian units of northern Sinai (including the Areif El Naqa) was given by Bachmann and Kuss submitted. Based on semiquantitative microfacies analysis the factors controlling deposition in the different systems tracts in northern Sinai during the Aptian-Cenomanian have been evaluated (Bachmann and Kuss submitted). The influence of 3rd order sea level oscillations on ramp deposition was reconstructed on the basis of facies patterns, sedimentary geometries, and the distribution of microfacies types. An important observation is that in the Aptian-Cenomanian of northern Sinai relative sea level changes do not result in a simple shift of facies belts but rather cause complex modifications of the lateral facies distribution by affecting biological productivity and the current systems (best visible in the microfacies distributions of the northerly exposed sections; Bachmann and Kuss submitted). Strong sea level drops result in exposure of the inner ramp or in the development of shallow restricted facies. In these horizons dolomites and/or pedogenetic overprints can be found. To understand the development of the prominent rudist biostromes during the HSTs, two factors have to be considered: Firstly, during constant sea level, the carbonate production rapidly reaches the catch-up stage and the steady progradation generates low-energy, open marine ramp environments. Secondly, rudist associations are highly adapted to muddy, low-energy environments, thus, their growth maximum on the Mid-Cretaceous northern Sinai ramp is reached during the HST.

A first description of the Turonian facies relations of Sinai and neighbouring areas was given by Kuss (1992b) and was extended to the Senonian interval in central east Sinai by Lüning et al. (submitted a). The models are to a great extent also valid for the Areif El Naqa area.

9.5.2. Hemipelagic deposits

A significant part of the sedimentary succession at the Gebel Areif El Naqa is represented by hemipelagic pelites of the middle neritic to upper bathyal. These deposits contain variable amounts of planktonic foraminifera and calcareous nannofossils. An important parameter represents the planktonic-benthonic (P/B) foraminiferal ratio. The P/B ratio was routinely determined for all samples by counting traverses with a minimum of 300 specimens in the unfractonated washed residue. The value is calculated in percent plankton or benthos in relation to the total number of counted foraminifera. At Areif El Naqa it was possible to determine P/B ratios for the upper Coniacian-Santonian, Maastrichtian and Paleocene-lower Eocene. Under certain conditions, the percentage of planktonic foraminifera in foraminiferal bottom assemblages increases with increasing water depth. The ratio is considered to depend on the relative difference between the productivity of planktonic species, which are in greater densities in open oceanic environments, and the productivity of benthonic foraminifera which is higher in neritic environments than in the deeper oceanic ones (Phleger, 1964; Reiss et al., 1974; Gibson, 1989; Van der Zwaan et al., 1990).

Detailed studies of the late Maastrichtian (Lüning et al., submitted c) and Paleocene (Lüning et al., submitted b) hemipelagites of eastern Sinai have shown that the Maastrichtian (and probably late Coniacian-Santonian) P/B fluctuations are related to surface water productivity changes while the Paleocene P/B cycles in general can be used for paleobathymetric reconstructions.

9.6. Depositional history

The discussion of the mid Cretaceous to early Eocene depositional history at Gebel Areif El Naqa is arranged after chronostratigraphic stages and lithological units (formations). Sequence boundary names refer to the
regional sequence stratigraphic models for the Albian-Cenomanian of northern Sinai (SB8-SB16; Bachmann & Kuss, submitted), the Turonian to Maastrichtian of central east Sinai (TuSin to Ma/DaSin; Lűning et al., submitted a) and the Paleocene of central east Sinai (DaSin-I to YpSin-I; Lűning et al., submitted b), provided a convincing correlation of these sea level drops has been established (Figs. 9-2, 9-3, 9-4). A detailed comparison with the regional schemes is presented in a later chapter of this contribution. Note that the frequency of the sea-level changes resolved varies within the studied Albian-Eocene interval. On the basis of the comparison with the geochronological time-scales of Gradstein et al. (1995) and Berggren et al. (1995) (Fig. 9-2), the cycles of the late Albian-Cenomanian are attributed to 3rd order, of the Turonian to Maastrichtian to 2nd-3rd order and of the Paleocene-early Eocene to 3rd order. A similar change from slow sea-level changes during the late Cretaceous to faster fluctuations during the Paleogene has been also noted by Mancini & Tew (1997) for the American Gulf Coastal Plain.

The present study is based on a biostratigraphic age model and does not attempt to provide an in-depth correlation of the existing lithostratigraphic schemes used in the region. Owing to the intermediate position between mainland Egypt and Israel, two different lithostratigraphic systems have been used for the Cretaceous-Paleogene in Sinai (Fig. 9-5). While the Egyptian formations are mainly based on Ghorab (1961), the Israeli nomenclature was developed by Flexer (1968) and Bartov et al. (1972). Correlations of the different schemes are included for example in Kerdany & Cherif (1990), Kuss (1992a) and Kora & Genedi (1995). Detailed lithological descriptions of the formations in central east Sinai can be, for example, found in Bartov et al. (1972), Ziko et al. (1993) and Kora & Genedi (1995).

### 9.6.1. Late Albian

The marine late Albian succession of the Areif El Naqa (section A10, Figs. 9-3, 9-6) disconformably overlies early Cretaceous alluvial sandstones. The best exposures are situated in the central part of the anticline along the lower parts of the E-W running escarpment of the northern flank (Fig. 9-1C). Guided by a section taken in a previous study further east and the columns given by Bartov et al. (1980) a new detailed stratigraphic log (A10, Fig. 9-6) with a total thickness of 130 m has been measured.

Predominant lithologies are marls, marly siltstones and limestones - the latter often dolomitized (early diagenetic). Within the lower portion, nearshore, often siliciclastically influenced sediments prevail that are overlain by dolomitic calcareous lithologies. The successions represent the southernmost marine exposures from northern Sinai. More complete Albian successions of nearly 300 m carbonates have been described by Bachmann et al. (1996) from outcrops further north. Sequence stratigraphic subdivisions allow to correlate them with the late Albian units of Areif El Naqa, based on biostratigraphic and facies considerations (Bachmann and Kuss submitted). All Albian carbonates of the northern Sinai were formed in shallow water environments, mainly within backshoal areas of a shallow, north-dipping ramp. The most prominent carbonates in the Areif El Naqa section represent four limestone beds which are characterized by abundant amounts of rudists. Sedimentary structures of three beds indicate biostromal growths, one yields rudist-buildups of up to 1.2 m length and 0.3 m height (Fig. 9-7C). Laterally, this bed shows a characteristic thickening and thinning, ranging from 0.6 to 1.8 m with major buildups in the more massive parts. While their top surface forms a nearly horizontal line (with mainly rudist-rudstones and wackestones, only subordinate mudstones), their base is formed by concave depressions, resulting in the laterally irregular undulations. Dolomitized rudist framestones and rudstones are the prevailing lithologies within these thicker parts of the bed. The massive limestones of the upper third are mainly composed of early diagenetic dolomites (plus late diagenetic overprintings). Only a few macroscopic textures allow the identification of an emersion horizon, followed by cross-bedded skeletal dolomites and a massive, rudist-bearing unit. However, the carbonates below are often composed of packstones and grainstones which hold oolite or ostracod accumulations (Fig. 9-8A), often with intercalated thin seams of angular to subrounded quartz grains. Wackestones and mudstones follow with minor frequencies, few of them with mainly badly preserved foraminiferal and algal remains of undeterminable miliolids and *Permcocalvus* sp., respectively (Fig. 9-8C).
### Stages

<table>
<thead>
<tr>
<th>Stage</th>
<th>Coniacian</th>
<th>Campanian</th>
<th>Maastrichtian</th>
<th>Paleocene</th>
<th>Eocene</th>
</tr>
</thead>
<tbody>
<tr>
<td>late</td>
<td>late</td>
<td>early</td>
<td>late</td>
<td>early</td>
<td>late</td>
</tr>
</tbody>
</table>

#### Fossil Zones
- *M. aragonensis*
- *M. formosa formosa*
- *M. subbotinae*
- *M. velascoensis*
- *G. pseudomenardii*
- *I. pusilla pusilla*
- *M. angulata*
- *P. uncinata*
- *P. trinidadensis*
- *P. pseudo-bulloides*

### Sequences
- NP12
- NP11
- NP10
- NP9
- NP7/8
- NP6
- NP5
- NP4
- NP3
- NP2

#### Relative Sea-Level History:
- Early rising
- Late falling

#### Syrian Arc Inversion
- Post-extensional phase
- Syrian Arc inversion (in pulses)
Fig. 9-3. Correlation of sequence boundaries in the late Albian-Turonian. Overview of sequence boundaries in Fig. 9-2. Location map in Fig. 9-1C, legend in Fig. 9-11. Lateral distance between sections not to scale.
Fig. 9-4. Correlation of sequence boundaries in the Coniacian to lower Eocene. Overview of sequence boundaries in Fig. 9-2. Curve with data points (black circles) shows the foraminiferal planktonic/benthonic ratio with 100% plankton (P) to the left and 100% benthos (B) to the right. Location map in Fig. 9-1C. Legend in Fig. 9-11. Lateral distance between sections not to scale.
The silty marls and limestones of the basal Late Albian succession are interpreted as having been deposited during rising sea level within a lagoonal setting. The sequence stratigraphic assignment is unclear (late TST or HST), in contrast to the overlying rudist bed which seems to represent the late HST. Strong dolomitization processes at its top have been attributed to an extreme shallowing which suggests the presence of a sequence boundary (SB8). Lagoonal marls and thin limestones above are overlain by another silty marl unit that co-occurs with biogenic limestone-marl couplets that suggest a deeper water environment of a shallow ramp. Within the lower parts, ammonite findings of *Knemiceras uhligi subcompresso* confirm a late Albian age of the marine succession. Discussions of the stratigraphic value of *Knemiceras* (including findings from northerly exposed localities) are given in Geyer et al. (in press). A nearly 45 m thick unit of marls, limestones and dolomites above is interpreted as having been deposited within lagoonal and tidal flat environments - including the rudist-bearing dolomites from the top, representing the uppermost part of a massive HST. Above SB9, similar lithologies appear within a LST, mainly of tidal flat origin. Above the dolomitic TST and HST sequence boundary SB11 was interpreted, based on regional comparisons with sections further north. Sediments of a sequence developed in other parts of northern Sinai (bounded by SB10 and SB11, Bachmann & Kuss, submitted) have not been found at the Areif en Naga. On top of the following LST, the Albian-Cenomanian boundary has been established, based on lithostratigraphic considerations.

### 9.6.2. Cenomanian

The exact position of the Albian-Cenomanian boundary in Sinai has been discussed by several authors (e.g. Bartov & Steinitz, 1978; Lewy, 1988; Said, 1990b; Kuss, 1992a). Because of the absence of indicative fossils, we follow Bartov et al. (1980) who used the marls of the ‘En Yorqeam Member’ to define the base of the Cenomanian based on lithostratigraphic comparisons with age-equivalent findings in Israel. The Cenomanian/Turonian boundary is clearly defined by deep water limestones with planctonic foraminifera and a characteristic association of lowermost Turonian ammonites, described by Bartov et al. (1980). The excellent exposures of the Cenomanian succession have been studied along the steeply north-dipping flanks of the anticline (north of the central depression and the E-W running escarpment - section A9, Figs. 9-1C, 9-9), where a carbonate/marl succession with a total thickness of 108 m was measured. Most of the limestones are of shallow-water origin, while the intercalated dolomites may be of early diagenetic origin. Moreover, major parts of the whole succession may be secondarily dolomitized, especially along the easterly exposures. Few carbonate/halite horizons show imprints of hardgrounds, nearshore/pedogenic overprint (Fig. 9-8D), rhizolithic structures (Fig. 9-7D) or sedimentary discontinuities. These observations are in good accordance with characean findings and ostracods in the overlying marl unit. Some of the ostracods indicate a brackish mode of living. In general, the marls hold poor assemblages of benthic foraminifera and often badly preserved ostracods.

Within the studied Cenomanian section, several limestones yield rich microfaunas. Besides few benthic foraminifera of *Chrysalidina gradata*, the rudist-bearing floatstones hold reworked or encrusted calcisponges of the genus *Sphaeractinia* (Fig. 9-8F), also found in age-equivalent rocks of Jordan (Kuss, 1992a). In most cases the rudists are of broken debris – often with micrit-infilled borings – indicating a reedimentation in rather quiet backshoal areas, where foraminiferal wackestones represent the most characteristic facies types. Here, often diverse assemblages of benthonic algae and foraminifera occur. Within the first group, *Sphaerulina schroederi* (Figs. 9-8I), *Pterocycladus* sp. (Fig. 9-8C) and *Aciculina* sp. are most frequent. Rhodophyceans are scarce and were found only in one layer with the pseydonellean alga *Pseudothallithium album* (Fig. 9-8F). According to Kuss & Conrad (1991) and Kuss (1994) this microflora is restricted to Cenomanian carbonates from age-equivalent strata of the region. Besides *C. gradata* the following foraminifera have been proven from several horizons: *Pseudoelminia dromimensis* (Fig. 9-8G), *Dicyclus* sp., *Hesyonema* sp., *Pyrgo* sp., *Nezuccata* sp., *N. convolita*, *Biconcava bentori*, *Vallumomina* sp., *Fischerina* sp. and *quinqueloculina miliolids*. *Praealveolina ochracea* (Fig. 9-8I) occurs in one horizon only. This foraminiferal assemblage characterizes Cenomanian limestones of the Tethys (Schroeder & Neumann 1985).

According to Bachmann and Kuss (submitted) the Cenomanian limestones were mainly formed in shallow backshoal environments of the southwestern prograding late Aptian-Cenomanian ramp. They are represented by wackestones, mudstones and few intercalated (mainly rudist-bearing) floatstones. In contrast to the underlying late Albian and the overlying Turonian carbonates, high energetic sediments like grainstones or packstones are of very subordinate importance. Sedimentation seems to keep pace with the gradually rising Albian-Cenomanian sea level and the shallowing intervals are probably due to minor sea level oscillations (Fig. 9-2). All Cenomanian sequence boundaries are interpreted as type 2. Lowstand systems tracts are partly missing. Above SB11 (about 4 m below the Albian-Cenomanian boundary) the ostracod-bearing dolomitic LST-limestones are overlain by

Fig. 9-5. Correlation of Egyptian and Israeli formations in eastern Sinai (modified after Kora & Genedi, 1995).
Fig. 9-6. Albian succession in section A10. Location map in Fig. 9-1C, legend in Fig. 9-11.
Fig. 9-7. Field photographs from Areif El Naqa. A Anticline viewed from west, relief is predominantly formed by hard Cenomanian-Turonian (often dolomitic) limestones. Beduin camps in foreground for scale. B Late Albian-earliest Cenomanian succession (110 m thick, section A10, Fig. 9-6, location map in Fig. 9-1C). Stippled line follows the unconformable contact between early Cretaceous continental sandstones and overlying late Albian carbonates. The arrow indicates the biostromal rudist horizon illustrated in (C) which is situated 8 m above the base of the marine late Albian strata. The Albian-Cenomanian boundary is developed in the upper part of the massive limestones / dolomites (cf. Bartov et al. 1980). C Closer view of the lower part of the late Albian succession with biostromal rudist limestones which show pronounced lateral thickness changes. Backpack at central right for scale (8 m above basal unconformity in section A10). D Early Cenomanian limestones of the late HST (near SB12). Pen indicates vertical structures interpreted as rhizolithic horizon. E Lower Turonian marls (80 m thick) with characteristic limestone bed in the middle, overlain by Upper Turonian dolomitic limestones (section A3, Fig. 9-3, location map in Fig. 9-1C). F Intra-chalk unconformity at the Santonian-Campanian boundary in section A6 (see Figs. 9-4, 9-11). G Slumped Campanian bedded chert above Santonian-lower Campanian hemipelagic chalk; slumping probably induced by tectonic uplift phase (1 km east of section A6, see Fig. 9-1C). H Intensely bioturbated medium- to coarse-grained sandstone from the upper Campanian-lower Maastrichtian. Boulders made of Eocene dolomite are obscuring part of the outcrop (section A1b, see Fig. 9-10).

Fig. 9-8. Typical microfacies elements from the Albian-Turonian units at Areif El Naqa. A Late Albian packstone, almost entirely composed of ostracodes with subordinate quartz grains, overlying micritic mudstone with a sharp erosional contact (early post SB14-LST, section A10, sample AN I 18). B Late Albian rudstone (lower half of picture) composed of densely packed gastropods and bivalve fragments, disconformably overlain by rudstones-grainstones that are mainly composed of reworked gastropods from below. Note the peloidal infillings of the shells in both parts (early pre-SB14 HST, section A10, sample AN I 9). C Late Albian wackestone with skeletal debris, including few algal remains (centre) of the udoteacean Permoactinina sp. (early pre-SB14 HST, section A10, sample AN I 6). D Early Cenomanian mudstone-wackestone with irregular voids, partly filled by micritic intraclast-bearing internal sediment. The original nearshore sediment has been overprinted by pedogenic processes (latest pre SB11 HST, section A9, sample AN Ila II). E Late Cenomanian wackestones with colonies of the peysonneliacean alga Pseudolothothamnium album (here: late post-SB16 LST, section A9, sample AN Ila 40). F Sphaeractinina-bearing rudstone with broken and subsequently bored rudist fragments. The calcisponges are frequent in the Cenomanian rudist biostromes (here of the late HST within the latest Late Albian sequence (pre-SB11, sample AN Ia 1). G Wackstone with several tests of Pseudolothothamnium album (late Cenomanian, post-SB11 TST, section A10, sample AN Ia 7). H High energy Turonian ostracod grainstone; note that the ostracod aggregates consist of several interlocked shells of different sizes containing residual micritic material which was protected from erosion by the surrounding shells (section A3, sample A3-10). I Wackestone-packstone with micritic intraclasts arranged in layers suggesting significant reworking (Upper Turonian, section A1, sample A1-7). J Early Cenomanian wackestone with skeletal molluscs (among them many gastropods), Suppiliulina schroederi (dasyycladacean), Praealveolina cretica and quinqueloculine miliolids (post-SB11 TST, section A10, sample AN I 10a). K Reworked oolithic wackestone-packstone with ostracods. Ooidal cores often formed by ostracod shells. Note sharp boundaries between micritic and bioclastic dominated material (Upper Turonian, section A1, sample A1-5). L Foraminiferal wackestone with abundant specimens of Discocina sp. (Turonian, section A1, sample A1-16A).
Fig. 9A. Cenomanian succession in section A9. Location map in Fig. 9C, legend in Fig. 9D.
wackestones with miliolids and few algal remains, interpreted as the early HST. The two-partitioned marts above are interrupted by a shallow marine limestone bed (mfs) containing a highly diverse microfauna and microflora. The upper marl with hedbergellids represent the late TST while dolomites with rhizolithic structures at the top form the HST. SB12 is interpreted to coincide with the shallowest part and is overlain by very similar facies types of the next LST. The first limestones with benthonic microfossils mark the onset of the next TST which also contains wackestones with rudist debris in the higher portions. A thick interval of marl-limestone intercalations above is interpreted as HST, where a subsequent shallowing of the cyclic limestones can be proven. At the top, stromatolitic laminates occur that coincide with SB14. The marts above hold brackish ostracods and characean oogonies which are attributed to the next LST. Its upper beds are more carbonatic and the first normal marine microfaunas are again taken as evidence for the beginning TST. During its late stages, reworked rudist fragments occur and gradually transit into dolomitic beds that are attributed to the HST. Green marts of probably brackish/fresh water origin have been interpreted to coincide with SB 16 (Figs. 9-2, 9-9). SB15 which has been reconstructed in age-equivalent sections further north (Fig. 9-2) (Bachmann & Kuss, submitted) was not proven at Areif El Naqa. The overlying LST with few miliolids and Pseudolithothamnium album grade into deeper water limestones with frequent planctonic foraminifera and ammonites at the top. The latter limestones mark the Cenomanian/Turonian boundary and are overlain by marts of sequence TuSin, described below.

9.6.3. Lower Turonian

The Lower Turonian at Gebel Areif El Naqa is characterized by mainly soft lithologies and is termed Abu Qada Formation (Ora Shales in the Israeli nomenclature). The succession can be tripartitioned into lower hemipelagic green marts, a massive (partly dolomitic) limestone bed with thickness regionally varying between 7.3-40 m (Bartov et al. 1980), and an upper unit consisting of siltstones, shales, bedded and nodular gypsum and micritic limestones. Rather abrupt transitions from marly to calcareous facies as well as variations in thickness occur in all three units of the Abu Qada Fm. (Bartov et al. 1980).

The Lower Turonian is developed typically in section A3a (Figs. 9-3, 9-7E). The presence of the planktonic foraminifera Whiteinella archaeoecretacea, Guembelitria cenomania, Heterohelix moremani and H. reussi in the lower hemipelagic green marts of this section suggests a biostratigraphic age of W. archaeoecretacea Zone or Helvetoglobotruncana helvetica Zone. In the same interval we found rare occurrences of the ostracods Spinoleberis yatavataensis and Paracypria triangularis which previously have been only described from the Turonian (Rosenfeld & Raab 1974). Bartov et al. (1980) described findings of early Turonian ammonites at the base [e.g. Mammites nodosoides (Schloth.), Neopychites cephaloites (Court.) and species of Fagesia and Thomasites] and the top (e.g. Choffaticeras luciae trisellatum Freund and Raab) of the green marl. The hemipelagic character of this unit is based on the presence of planktonic foraminifera. We interpret the lower green marts as a transgressive systems tract (pre-TuSin, Figs. 9-2, 9-3).

The lower part of the overlying hard micritic limestone in section A3a is massive and partly bioturbated with gastropods while the upper part is composed of dolomitic, partly graded grainstones and packstones containing mainly rudists, gastropods and other molluscan shells. The presence of these inner shelf carbonates indicates a progressive shallowing in relation to the underlying hemipelagites so that the limestone bed is interpreted as a highstand systems tract (pre-TuSin).

In section A3a, the massive limestone is directly overlain by siltstones, shales and bedded gypsum of peritidal-supertidal character which point to a pronounced shallowing event. We interpret the limestone-siltstone boundary as sequence boundary (TuSin). Restricted shallow marine to hypersaline conditions are also indicated for this lowstand interval by the abundant occurrence of the ostracod Neopychites vandenboldi in section A3a, and the bivalves Cerithium sp. and Caryocorbula sp. which have been described by Bartov et al. (1980) from this horizon at the Areif El Naqa. A similar, age-equivalent lithology and fauna was also described from the anticlines in the Negev Desert by Braun et al. (1987) who also interpreted a pronounced shallowing for this interval.

Development of higher order paleobathymetric cycles within the post TuSin LST in section A3a may be inferred from several horizons of reworked lagoonal ostracod-rich carbonates and marts which are intercalated into 3-4 m thick sabkha-type gypsum deposits. A very similar facies and sequence stratigraphic development for the Lower Turonian as at Areif El Naqa has been also reconstructed for central east Sinai (Lüning et al., submitted a).

9.6.4. Upper Turonian

The transgressive surface in section A3a is interpreted to lie between the uppermost gypsum bed and overlying lagoonal pelites (Figs. 9-2, 9-3). While the lower part of the TST is still lying in the upper part of the Abu Qada Formation (Ora Shales), the lagoonal pelites evolve upwards into harder, partly dolomitic inner shelf limestones and calcareous marts which partly contain flint nodules. The hard carbonates are grouped into the Upper Turonian-Lower Coniacian Wata (Gerofit) Formation and are developed in section A1a (Fig. 9-4) (and probably A3b). Occasionally, peritidal laminates (partly with dewatering structures) are intercalated into the inner shelf facies, possibly indicating the development of higher frequency paleobathymetric cycles. The inner shelf carbonates can be subdivided into a low and a higher energetic facies. Further subdivisions of these facies types
have not been attempted for the Upper Turonian because of the high potential for autodynamic-autocyclical sedimentary processes leading to an unpredictable lateral shift of the different microfacies patches as evidenced by frequent reworked fabrics (Figs. 9-8H, 9-8L, 9-8K). This is also supported by data from Bartov et al. (1980) who observed lateral lithostratigraphic variations which cannot be related to the local anticlinal structure. A more detailed description and discussion of this matter for central east Sinai where a similar facies in the Upper Turonian is developed is included in Lüning et al. (submitted a).

The Upper Turonian-Lower Coniacian low energetic inner shelf carbonate facies consists of calcareous marls, mudstones and wackestones containing variable amounts of peloids, ostracods, miliolids, other benthonic foraminifera (Fig. 9-8L), gastropods, oysters, other molluscan shells, bryozoa and echinoids. Sedimentary structures are in general missing.

The higher energetic inner shelf carbonate facies of this interval consists of grainstones, rudstones, packstones and wackestones mainly composed of reworked components of the low energetic carbonate facies. Additional components are superficial ooids and rudist debris. Section A1a (Fig. 9-3) includes a 4 m thick interval with rudist in life position. A bioclastic packstone in the uppermost part of section A1a holds the Udoteacean alga Boueina pygmaea (Pia) which from northern Sinai (Gebel El Minsherah) was previously described by Kuss (1994). While in some cases a large part of the mieritic matrix is still preserved but clearly reworked, relictic micritic OC layers (Fig. 9-8K) and rip-up clasts (Fig. 9-8I) can be observed in other examples. Typical sedimentary structures of this facies include crossbedding, grading and channels.

Planktonic foraminifera are not present in the Upper Turonian at Areif El Naqa which has been previously recognized by Ayyad et al. (1996). In section A1a the ostracod Neocyprideis vandenboulli has been found which according to Hataba & Ammar (1990) has its last occurrence in Egypt in the Upper Turonian (Wata Fm.). The Upper Turonian carbonates of the Wata Fm. at Gebel Areif El Naqa are interpreted as post TuSin TST deposits.

9.6.5. Coniacian

The overlying Coniacian of the Themed Formation (Zihor Fm., according to the Israeli nomenclature) and therefore the upward continuation of the genetic sequence has not been found properly exposed in section A1a. However, a detailed description from other locations at the Areif El Naqa structure was previously given by Bartov et al. (1980). The lower part of the succession is reported to consist mainly of "open marine" chalky limestone and marls while the upper part of the unit is dominated by bioclastic limestones of inner shelf type. An early Coniacian age was assigned to the open marine part of the Coniacian succession through the ammonites Roemeroceras parnesi Levy and Barroisiceras onilahyense Basse whereas the inner shelf carbonates have been dated as late Coniacian since they contain Roemeroceras tunisiense (Hyatt) (Bartov et al., 1980). We interpret the open marine deposits as the late post-TuSin TST and the overlying bioclastic limestones as undifferentiated HST and post CoSin-LST (Fig. 9-2). The slightly younger age estimation for the post-TuSin TST at Areif El Naqa in contrast to central east Sinai (late Turonian, Lünig et al., submitted a) may be explained by correlation problems between biostratigraphic data from microfossils and ammonites.

While in many parts of central east Sinai the associated sequence boundary CoSin can be traced near the base of a siliciclastic succession (Lünig et al., submitted a), in the basinward-lying carbonate facies of the Areif El Naqa area (Lewy, 1975) only the "correlative conformity" (Vail et al., 1977) of this boundary seems to be developed so that a distinct horizon for the sequence boundary cannot be given for the study area. Bartov et al. (1980) report strong lateral variations in lithology and faunal composition for the Coniacian unit at Gebel Areif El Naqa. A Coniacian section from the Areif El Naqa was also described by Khalifa & Eid (1995).

9.6.6. Upper Coniacian - Santonian

The Upper Coniacian - Santonian deposits in the Areif El Naqa area are composed mainly of hemipelagic chalks which often contain moderate to abundant assemblages of benthonic and planktonic foraminifera, ostracods and calcareous nannofossils. Bartov et al. (1980) described the unit as Menuha Fm., whereas Egyptian workers include the interval sometimes into the upper Themed / Matulla Fm. (e.g. Ziko et al., 1993) and sometimes into the lower Sudr Chalk (Allam & Khalil 1988). We interprete the contact between the inner shelf-type bioclastic limestones and the overlying hemipelagic chalk unit as a transgressive surface (Figs. 9-2, 9-4). We investigated these TST deposits at the northern (section A1b; Figs. 9-1C, 9-10) and at the southern flank (section A6; Figs. 9-1C, 9-11) of the anticline. Planktonic foraminifera (Fig. 9-12) indicate a biostratigraphic age of Dicarinella concavata and D. asymetrica Zones and calcareous nannofossils mark the biotones of Lucianohabds muleiformis, Reinhardtides anphphorus and L. cayeuxii. Additional biostratigraphic evidence for the Upper Coniacian-Santonian age of the succession comes from ostracods in section A1b which for example include Cythereis diversereticulata (Coniacian-Maastrichtian), Brachycythere angulara (Coniacian-Maastrichtian), Cythereis cretaria diversereticulata (Santonian-Campanian), Cristaleberis reticulata (Coniacian-Paleocene), Cythereis jordanensis (late Santonian-early Campanian) and Cythereis mesa mesa (late Santonian-Paleocene) (ranges described in the literature so far given in parantheses).
Fig. 9-11. Range chart of planktonic foraminifera and semiquantitative distribution chart of calcareous nannofossils in section A6. Legend for nannofossil abundances in Fig. 9-13B: legend for biozones in Fig. 9-4. Curve with data points (black circles) shows the foraminiferal planktonic/benthonic ratio with 100% plankton to the left and 100% benthos to the right. Location map in Fig. 9-1C.
Fig. 9-12. Planktonic foraminifera from the Upper Cretaceous of Areif El Naqa [except specimens 3, 20/21, 22/24 which come from sections central east Sinai (Lüning et al., submitted a)]. Bar represents 300 μm.

1-2  *Whiteinella archaeocretacea*, lower Santonian of section A6 (sample A6-5).

3-5  *Archaeoglobigerina bosquensis*, specimen 3 from upper Coniacian/lower Santonian of section D (sample D1-3), specimen 4/5 from the lower Santonian of section A6 (sample A6-1).

6-7  *Whiteinella baltica*, Santonian of section A6 (sample A6-6).

8-10  *Dicarinella asymetrica*, Santonian of section A6, specimen 8 from sample A6-8, specimen 9/10 from sample A6-1.

11-13  *Hastigerinoides subdigitata*, Santonian of section A6, both specimens 11 and 12/13 from sample A6-14.

14-15  *Globotruncana bulloides*, lower Campanian of section A6 (sample A6-17).

16-18  *Globotruncanita elevata*, lower Campanian of section A6, specimen 16/18 from sample A6-19, specimen 17 from sample A6-17.

19-21  *Globotruncanana arca*, specimen 19 from the lower Campanian of section A6 (sample A6-17), specimen 20/21 from the Campanian of section T2 (sample T2-37).

22-24  *Rosita plummerae*, specimen 22/24 from the Campanian of section T2 (sample T2-23), specimen 23 from the lower Campanian of section A6 (sample A6-17).


28-30  *Plummerita reicheli* [sensu Masters 1993], uppermost Maastrichtian, both specimens 28/30 and 29 from sample A5-17.
We interpret that the maximum flooding surface in section A1b (Figs. 9-2, 9-4, 9-10) lies below a pebbly glauconitic dolomite indicating a first higher order sea-level fall which, however, is followed again by a short-lasting hemipelagic marly episode characterized by significant amounts of planktonic foraminifera. The reconstructed mfs lies within the L. cayeuxii Zone which corresponds well with the findings in central east Sinai (Lüning et al., submitted a).

9.6.7. Campanian - Maastrichtian

The Campanian succession is composed of chalks, flints, bioclastic limestones and sandstones (sections A1b, A6; Figs. 9-10, 9-11) and is grouped into the lower Sudr Chalk (Mishash Fm.). In section A6 the Santonian-Campanian boundary is developed as an intra-chalk unconformity marked by reworked marl clasts and phosphoritic pebbles (Fig. 9-7F). The horizon is interpreted as sequence boundary Sa/CaSin (Fig. 9-2). The chalk below the unconformity has a biostratigraphic age of Discinella asymetrica Zone (Fig. 9-12; 8-10) and the overlying chalk has been attributed to the Globotruncanita elevata Zone (Braasiona parca Zone by nanofossils). Age-equivalent sea-level falls have been also described by Lewy (1990) from Israel and Kuss & Malchus (1989) from the Eastern Desert. The lower Campanian chalk in section A6 is overlain by a 10 m thick unit of chalky flint (early Campanian B. parca Zone) which in a similar lithology and thickness is also developed in section A1b. In the latter section the unit contains abundant accumulations of silicified oysters which Bartov et al. (1980) and Bartov & Steinitz (1982) interpreted as oyster bioherms. In section A6 the unit is partly folded with wavelengths ranging around 10 m (Fig. 9-7G). The folds are mostly asymmetric and the folded horizon is detached from a slightly undulated lower chalk substratum. The flint deposits are overlain by medium to coarse grained sandstones which are composed predominantly of quartz and contain only a few sub-angular mm-scale micritic clasts, are well sorted, have a relatively high porosity and are cemented by sparite. While the sandstones in section A1b (Figs. 9-7H, 9-10) reach a thickness of about 15 meters, the same horizon in section A6 (Fig. 9-11) is restricted to 10 centimeters (Fig. 9-4). A most probably age-equivalent glauconitic marly fine sandstone with a thickness of about 1 m is also developed in section A2 (Fig. 9-4) and is overlain by dolomitic chalks and bioclastic carbonates of undifferentiable Maastrichtian age (several Rugoglobigerina species found, however no biostratigraphic index forms). Other locations at Areif El Naqa with the development of similar sandstones and conglomerates have been described in detail by Bartov et al. (1980). In section A1b (Fig. 9-10) the siliciclastics fine upwards into fine sandstones, siltstones, silty marls and are overlain by 10 m of hemipelagic marls of the late Maastrichtian (Abathomphalus mayaroensis Zone by foraminifera / Micula murus Zone by nanofossils) with high percentages of planktonic foraminifera. Only three kilometers to the east in section A5 thickness of the same biostratigraphic interval has tripled to 30 m. A few kilometers north of the anticline in section A7 (Figs. 9-9C, 9-13B), a hemipelagic chalk succession covering almost the whole Maastrichtian from the upper G. falsostuartii Zone to the A. mayaroensis Zone (Arkhangelskiella cymbiformis to M. murus Zone by nanofossils) is developed. Thickness of the A. mayaroensis / M. murus Zone here is comparable to that in section A5 (Fig. 9-4). On the southern side of Areif El Naqa in section A6 (Fig. 9-10), the thin siliciclastic horizon is overlain by sandy chalks with an early Maastrichtian age of A. cymbiformis Zone.

In the sections A5 and A7 (Figs. 9-4, 9-13B) the studied Maastrichtian hemipelagites show more or less pronounced fluctuations of the planktonic-benthonic foraminiferal ratios which according to semiquantitative investigations of calcareous nanofossils and benthonic foraminifera are related to changes in surface water productivity (Lüning et al., submitted c).

9.6.8. Paleocene-Lower Eocene

The Paleocene to Lower Eocene part of the succession is termed Esna Shale (Taqyiye Fm. in the Israeli nomenclature) and has been studied in sections A1b, A5, A4, A7, A8 and W (Figs. 9-4, 9-10, 9-13A). The unit is predominantly composed of greenish hemipelagic marls with abundant planktonic foraminifera and calcareous nanofossils which allow a detailed biostratigraphic subdivision of the interval [range chart for section A8 in Marzouk and Lüning (in press: Fig. 7)]. Nearly all Paleocene biozones of planktonic foraminifera and calcareous nanofossils occur in the Areif El Naqa area, although not included in a single continuous section. While the most complete Paleocene-lower Eocene successions have been found in section A8 north of the anticlinal structure and in section W south of Areif El Naqa, in sections A1b and A5 which are located at the northern flank of the structure, only the biozones NP2-NP3 are developed, bounded below and above by significant unconformities (Figs. 9-1C, 9-4). In the latter sections, the Danian marls contain low amounts of subrounded to angular quartz grains of medium sand size until the lower P. uncinata Zone.

The Cretaceous-Tertiary boundary interval is exposed in sections A1b, A5, A7 and A8 and is characterized by a hiatus covering NP1 (P. eugubina and lower part of P. pseudobulloides Zone) and locally the Micula prinsii Zone (uppermost A. mayaroensis Zone). Only a few 100 m from section A8 in section A7, however, the biozone NP1 / lowermost P. pseudobulloides Zone has been proven. We interpret the horizon just above the K/T boundary at Areif El Naqa as being associated with a sea level fall and interpret it as a sequence boundary...
Fig. 9-13. Range chart of planktonic foraminifera and semiquantitative distribution chart of calcareous nannofossils. A Section W. B Section A7. For further explanations see Fig. 9-11.
(DaSin-1). The planktonic-benthonic foraminiferal ratio in the studied sections at Areif El Naqa shows a typical pattern which is also developed in central east Sinai. For the Paleocene of central east Sinai it has been shown that the major trends in the planktonic values can be used as a proxy for sea level changes (Lüning et al. submitted b). Many of the eustatic sea level changes reconstructed for central east Sinai (Fig. 9-2) are also documented in the P/B curves from the Areif El Naqa region. Among the less pronounced sea level events are the drops in planktonic values within the P. trinitatisensis Zone (NP2-NP3 boundary) of section A8 (sequence boundary DaSin-3 in Lüning et al., submitted b) and within the M. angulata Zone (NP4) of sections A8 and W (sequence boundary ThSin-1 in Lüning et al., submitted b). A significant sea-level drop is documented in sections A8 and W by a strong drop in planktonic values around the Igorina pusilla pusilla - G. pseudomenardii boundary (within NP5) which can be also observed in central east Sinai (ThSin-2 in Lüning et al., submitted b).

Furthermore, most of the other sequence boundaries developed during the G. pseudomenardii Zone in central east Sinai are resolved at Areif El Naqa. While sequence boundary ThSin-3 is marked by a hiatus in section A8 with NP6 missing, ThSin-4 is developed in sections A4, A8 and W within NP7/8 at the base of a hard chalk or micritic limestone with a thickness of 1-5 m. We interpret the bed as a firmground or hardground associated with intensified bottom currents during low sea level (see discussion in Lüning et al., submitted b). Sequence boundary ThSin-5 (Figs. 9-2, 9-4) at the NP7/8-NP9 boundary is not properly resolved at Gebel Areif El Naqa and may be amalgamated with ThSin-4 in section W. A pronounced hiatus with NP9 and the M. velascoensis Zone missing is developed in section A8 and its origin is discussed further below. In sections A4 and W, the M. velascoensis and M. edgari Zones are characterized by marls with high planktonic values indicating a pronounced deepening. In section A4 the lowermost Eocene hemipelagites contain significant amounts of fine quartz grains.

The soft hemipelagic succession is overlain by hard dolomites (often with flint nodules) of the Middle Eocene Thebes Formation. While in the sections in some distance to the anticlinal core (A4, A8, W) the dolomites overlie marls with a biostratigraphic age of M. edgari Zone / NP10, in more proximal locations (A1b, A5) the uppermost marls are dated as M. uncinata Zone / NP3. From analogy with interpretations for central east Sinai and other regions (see discussion in Lüning et al. submitted b) we interpret that deposition of the dolomites mark a pronounced shallowing event. In section A5 the basal part of the Thebes Fm. contains abundant glauconite which is probably associated with mechanisms leading to a current-induced firmground / hardground situation.

### 9.7. Comparison with the depositional history on the stable shelf and reconstruction of the uplift history at Gebel Areif El Naqa

In order to differentiate between (at least partly 'eustatic') regional and local, tectonically-induced sea-level changes in the Late Cretaceous-Paleocene, we compare our findings complemented by observations of Bartov et al. (1980) from Gebel Areif El Naqa with the depositional history on the tectonically quiet stable shelf in central east Sinai (Lüning et al., submitted a, b) (Fig. 9-2). A similar method was already employed in an early foraminiferal benchmark paper by Said & Kenawy (1956) who compared a synclinal and an anticlinal section (Nakhil and Gebel Giddi) from northern Sinai. It has to be pointed out that post-Jurassic sediments are not resolved at Gebel Areif El Naqa east Sinai (Liining et al., submitted a, b) (Fig. 9-2). A similar method was already employed in an early foraminiferal benchmark paper by Said & Kenawy (1956) who compared a synclinal and an anticlinal section (Nakhil and Gebel Giddi) from northern Sinai. It has to be pointed out that post-Jurassic sediments are not resolved at Gebel Areif El Naqa east Sinai (Liining et al., submitted a, b) (Fig. 9-2).

#### 9.7.1. Pre-late Albian

The pre-Albian development of the Areif El Naqa anticline which was not subject of this study was previously described by Bartov et al. (1980). According to these authors, first tectonic movements in the area of the Areif El Naqa are dated from the late Triassic / early Jurassic and seem to be a consequence of regional graben formation associated with the separation of the Turkish-Apulian terrane from Northeast Africa. Laterally differential subsidence, bounded by faults and flexures, as well as local folding are reported to have led to partial exposure, non-deposition and minor erosion. During the late Jurassic / early Cretaceous the area is interpreted to have been tilted slightly to the NW which locally has led to erosion (Bartov et al. 1980).

#### 9.7.2. Late Albian-Cenomanian

In mid Cretaceous times a gradually north-dipping carbonate ramp with southward retrograding facies belts had been established in northern Sinai (Bachmann et al., 1996). During the late Albian, Areif El Naqa was situated near the southern coastline of that ramp, which during the Cenomanian shifted further southwards. While the upper Albian sediments from Areif El Naqa differ lithologically from those of northern Sinai, mainly due to
shallow water depth, increased siliciclastic input and different biotic composition, the Cenomanian succession shows very similar lithologies and biota in both areas, associated with the overall transgression during this time. The Albian-Cenomanian sea level changes reconstructed for Areif El Naqa in general correspond to the sequence stratigraphic development in many other locations in northern Sinai (Bachmann & Kuss, submitted) indicating tectonic quiescence for Areif El Naqa during this period. This is also supported by observations of Bartov and Steinitz (1977) and Bartov et al. (1980) who report a rather constant thickness distribution for the Albian-Cenomanian Hazera Fm. throughout the north-central Sinai/Negev area. It may only be speculated whether deposition in northern Sinai during the Albian was still influenced to some extent by the relief formed by the mid Mesozoic extensional phase. A comparable scenario is assumed by Keeley (1994: 736) and Ayyad & Darwish (1996) for the early Cretaceous of northern Egypt. Obviously, the footwall areas in such a case would experience greater subsidence rates than the hangingwall areas which would show its effects in the lateral facies distribution. Nevertheless, to address this aspect properly, an areally higher resolution data base including subsurface data collected in northern Sinai by the oil industry is required.

9.7.3. Turonian-mid Maastrichtian

The Turonian to early Campanian sea level history in the studied sections at Areif El Naqa (with data from Bartov et al., 1980 for the Coniacian) corresponds well with the sequence stratigraphic development reconstructed for central east Sinai. Nevertheless, detailed mapping at the anticline by Bartov et al. (1980) showed the existence of more or less pronounced lateral facies and thickness changes for this stratigraphic interval. While in the Turonian these changes are reported to show no systematic trends related to the present anticlinal structure, the Coniacian and Santonian sedimentary patterns are to some extent linked to the anticlinal geometry (Fig. 9-2). Consequently, the lateral thickness and facies fluctuations in the Turonian Abu Qada Formation (Ora Shales) and the Wata (Geroft) Formation at Areif El Naqa are interpreted to be predominantly associated with autodynamic sedimentary processes suggesting the absence of major tectonic uplift movements at Gebel Areif El Naqa during the Turonian (Fig. 9-14A). Based on the isopach and facies distribution by Bartov et al. (1980), uplift seems to have started during the Coniacian (Fig. 9-14B), although regional sea level changes still have shaped the Coniacian-Early Campanian vertical facies development markedly as the overall good correspondence with the sea level history in central east Sinai (Fig. 9-2) (Lüning et al. submitted a) during this period suggests.

An important marker for an intensified period of uplift represent the mid Campanian to early/mid Maastrichtian siliciclastics developed in several sections which exhibit great differences in thickness and grain sizes (Figs. 9-4, 9-7H, 9-10). Biostratigraphically the stratigraphic range of that unit can be narrowed down to the interval G. ventricosa - G. gansseri Zone (post-B. parva to pre-A. cymbiformis Zone). In section A6 the siliciclastic horizon is underlain by folded cherts (see above and Fig. 9-7G) which we interpret as syntectonic slump folds rather than higher order tectonic folds because (a) asymmetry corresponds well with the orientation of the paleoslope, (b) similar small-scale fold structures in this lithological unit have not been observed in other places at Areif El Naqa, (c) underlying and overlying strata are less or not deformed and (d) the overlying sandstone suggests uplift and the presence of a significant relief during this time (see following discussion).

Because similar age-equivalent siliciclastics are not developed in other parts of Sinai (see Lüning et al. submitted a) and the Areif El Naqa anticline is interpreted to have been an isolated paleohigh surrounded by calcareous outer to inner shelf deposits, the sandstones and conglomerates must have been eroded from older strata exposed in the anticlinal core during that time (Fig. 9-14C). Recycling of the siliciclastics is also suggested by the mature composition of the siliciclastics because the potential distance for sorting and differential physico-chemical weathering processes is much too small to account for such mature siliciclastics. A similar reworking mechanism has been previously postulated by Bartov et al. (1980) who described conglomerates of this unit consisting of quartztic pebbles which they interpret to have been reworked from the underlying 'Arod Conglomerate' of early Cretaceous age. The thickest sandstone deposit with 15 m is developed at the northern flank in the area of section A1b (Figs. 9-1C, 9-10; see also section 3 in Fig. 10 in Bartov et al., 1980). The siliciclastics at the southern flank of the anticline are in general significantly thinner, especially in section A6 with 10 cm, or are missing at all. We assume that the maximum thickness in the north is linked to the asymmetrical structure of Areif El Naqa with an areally extensive, relatively low gradient northern flank which collects most of the eroded material in channels and a steeper southern flank which has a significantly reduced slope width (Fig. 9-14C). A comparison with the sequence stratigraphic history of central east Sinai shows that the period of tectonically induced siliciclastic sedimentation coincides with two regionally developed sea level lowstands during the mid Campanian (G. ventricosa Zone CaSin, Fig. 9-2) and around the Campanian-Maastrichtian boundary (Ca/MaSin, Fig. 9-2) which may have facilitated erosion and deposition of the siliciclastics.

9.7.4. Mid Maastrichtian-early Eocene

The siliciclastics are overlain by hemipelagites often characterized by high percentages of planktonic foraminifera which are typically developed also in other parts of Sinai. This suggests that late Maastrichtian
Fig. 9-14. Late Cretaceous-Early Tertiary syndepositional uplift history at Areif El Naqa. Position of studied sections is shown schematically. Vertical and horizontal dimensions not to scale. A No paleostructures are known from the area until the Late Turonian. B Uplift at Areif El Naqa is inferred to have started during the early Coniacian as evidenced by isopach data (Bartov et al., 1980). C During the late Campanian-early Maastrichtian lower Cretaceous sandstones became exposed in the anticlinal core, were eroded and redeposited at the flanks of the anticlinal structure. D Renewed uplift during the middle and late Paleocene resulted in a hiatus between the lower Paleocene and middle Eocene in the proximal sections (A1b, A5). See text for further details.
deposition at Areif El Naqa was again controlled predominantly by the (inter-) regional sea level development rather than by local tectonic movements. Maastrichtian facies distribution in sections A1b, A6 and A7 (Figs. 9-1C, 9-4) is directly connected to the paleostructure. Following the siliciclastic episode, hemipelagic conditions were proven in the northern (A7) and southern (A6) basinal sections already during the early Maastrichtian (A. cymbiformis Zone) while at the northern anticlinal flank (A1b) this occurred significantly later during the late Maastrichtian (A. mayaroensis / M. murus Zones). In addition, the lower Maastrichtian sediments have a more hemipelagic character in the distal northern section A7 than in the section A6 (which is closer to the anticlinal core) as evidenced by differences in planktonic values and siliciclastic influence (Figs. 9-1C, 9-4). The chalks in Section A7 contain abundant planktic foraminifera and only little or no quartz grains whereas the chalks in section A6 are sandy and are dominated by benthonic foraminifera.

The regional sequence boundary (DaSin-1) just above the K/T boundary is also developed at Areif El Naqa but seems to be enhanced by the inactive anticlinal relief or active uplift. At Areif El Naqa, the early Paleocene biozone NP1 is almost completely missing so that the associated hiatus is significantly greater than in central east Sinai. However, slightly outside the anticline to the north, the interval of NP1 / lower P. pseudohullioloides Zone is properly developed, supporting the local character of the prolonged hiatus. Small amounts of quartz grains within the hemipelagic deposits of NP2 and NP3 in sections A1b and A5 suggest that the siliciclastic source in the anticlinal core had been still active during this time. However, high sea level may have reduced the potential catchment area and, in addition, may have trapped most of the eroded detritus further upslope.

In the sections proximal to the anticlinal core (A1b, A5), the middle and upper Paleocene marls are missing and the P. uncinata Zone (NP3) is directly overlain by middle Eocene dolomites of the Thebes Formation (Fig. 9-4). The nature of the marl-dolomite contact cannot be clearly identified as tectonic thrust or unconformably sedimentary because of the soft, rather homogeneous appearance of the marls. Nevertheless, we favor the model of a sedimentary unconformity related to mid / late Paleocene uplift in combination with significant eustatic sea level falls (ThSin-1 to YpSin-1 in Lüning et al. submitted b) (Fig. 9-14D). The sections lying in a more distal position to the anticlinal core (A4, A8, W: Fig. 9-1C) contain rather undisturbed, complete Paleocene successions in which the typical (eustatic) sea level pattern as reconstructed for central east Sinai (Lüning et al. submitted b) is clearly developed (Fig. 9-4). The differences between the proximal and distal sections in relation to the anticlinal core suggest the presence of a pronounced paleorelief during this time. This corresponds well to the model of Cohen et al. (1990) who postulated the development of fault-fold sag basins adjacent to the steep flanks of the Syrian Arc anticlines.

In the distal northern section A8 (Fig. 9-4), a pronounced hiatus which covers the M. velascoensis Zone (NP9) has been found while the same interval is present in sections A4 and W. In central east Sinai the M. velascoensis Zone often appears unusually thin and partly is also absent. A convincing explanation for the locally and regionally inconsistent distributional pattern of this biozone has not been found. Possible explanations include repeated strong eustatic sea level falls as postulated by Haq et al. (1987), tectonic processes or intensified bottom-currents.

A marly fine sandstone with a typical rich hemipelagic planktonic fauna has been found in the early Eocene (Mi. edgari Zone / NP10) in section A4 (Fig. 9-4) which we interpret as a (turbiditic) mass flow deposit consisting of eroded siliciclastic material from the anticlinal core. In neighbouring sections (A8, W) age-equivalent horizons are devoid of quartz grains so that renewed uplift cannot be reliably interpreted from this siliciclastic deposit. According to Bartov et al. (1980) a terminal folding phase at Areif El Naqa occurred after the late Eocene. Summing up, we interpret that the Areif El Naqa area has remained tectonically rather quiet until the beginning of the Coniacian and experienced major uplift phases during the Coniacian (?-Santonian), late Campanian-early Maastrichtian and mid-upper Paleocene. Additional tectonic activity phases may have occurred at other times of the studied period, however may be masked by high sea level.

9.8. Comparison with the deformational history in other parts of the Syrian Arc

In the following we compare the uplift phases reconstructed for the Areif El Naqa anticline with the tectonic development in other parts of the Syrian Arc Foldbelt.

9.8.1. Pre-Coniacian

Until the end of the Turonian, Cretaceous lithologies in Sinai are in general uniform over wide areas with only gradual thickness changes (Bartov & Steinitz, 1977). As described earlier, facies and thickness changes developed at Areif El Naqa are attributed to autodynamic sedimentary processes. This suggests that the Syrian Arc has remained tectonically rather quiet in Northern Sinai during the post-extensional / pre-compressional pre-Coniacian.
9.8.2. Coniacian-Santonian

A first major uplift phase at Areif El Naqa is interpreted for the Coniacian-Santonian based on lateral facies and thickness changes which are linked to the anticlinal geometry (see above). Tectonic activity during this period was previously described from several other locations in the Syrian Arc. In Northern Sinai, for example, Upper Coniacian sediments are reported to be missing which was interpreted by Lewy (1975) as an uplift of northern Sinai during the late Coniacian while the central part of Sinai was still covered by the sea. At some locations in northern Sinai, erosion reached as deep as into Turonian horizons (Kerdany and Cherif 1990). Nevertheless, it has to be pointed out that the Coniacian erosion may at least in part be associated with an inter-regional and probably eustatic mid Coniacian sea level lowstand (CoSin; see above and Lüning et al., submitted a).

Based on subsidence curves derived from drill holes in Israel, Hirsch et al. (1995b) reconstructed several compressional pulses for the Negev Foldbelt from the Coniacian to lower Miocene with a clear differentiation into initial, main and late phases. They point out that deformational intensity within the Syrian Arc shifted markedly in time and space. Shahar (1994) subdivided the deformational history of the Sinai-Negev Foldbelt into three phases. On the basis of isopach and lithological maps he reconstructed a phase of simple folding for the late Turonian to middle Eocene which produced 30-50% of the up to 1200 m total amplitude of the different structures. The late Cretaceous uplift in Northern Egypt is also documented, although biostratigraphically poorly constrained, on seismic profiles where the upper Cretaceous to Oligo-Miocene sediments show clear onlap against the synsedimentary rising anticlinal structures (Ayyad & Darwish, 1996). Similar onlap structures have been found in seismic lines in the Palmyride Foldbelt (Chaimov et al., 1992).

In the Negev Desert, Braun et al. (1987) and Honigstein et al. (1988) reconstructed a major tectonic phase for the late Turonian to lower Santonian based on short-distance thickness changes in the upper part of the Turonian Nezer Formation, good correspondence between thick intervals and structurally deeper-lying areas, and biostratigraphically well constrained onlap patterns. From a section in the Giva’at Mador area in the northern Negev Desert, Zur et al. (1995) described two unconformities from the late Turonian to Santonian interval which they correlate with compressional phases. The earliest erosional unconformity truncates the top of the Upper Turonian-Lower Coniacian Nezer Formation and the lower overlying Santonian Menna Fm. A second erosional unconformity truncates partly or completely the Santonian Menna Fm. and is associated with a thin basal conglomerate (Zur et al., 1995). Regional sea-level fall during the mid Coniacian (sequence boundary CoSin) and at the Santonian-Campanian boundary (Sa/CaSin) may have also contributed to the formation of these unconformities. On the basis of isopach data, Steinitz (1976) reconstructed a ‘mild tectonic regime’ for most of the Coniacian to Santonian paleorelief was mainly formed prior to and during the deposition of the lower parts of the succession. This corresponds to our observation at Areif El Naqa, that besides tectonic uplift, the regional sea-level signal still influenced deposition during this period markedly. For the Judean Mountains in Israel, Flexer et al. (1989) postulated an early tectonic phase of early Late Coniacian age.

From the offshore area west of Israel, Mart (1994) described early Senonian folding of the Litani anticline. However, shortly after formation of the folding structure, still in the Senonian, extensional processes are reported to have started and resulted in fast subsidence in the Ptolemaïs-Basin along a tranform fault which had been formed during the late Triassic-early Jurassic rifting in the Eastern Mediterranean region. Mart (1994) explains the fast shift from compression to extension by complex collisional and subduction processes in the Afro-Eurasian collisional zone. He assumes that if the process is dominated by continent-continent or continent-terrane collision, a compressive stress field is developed in the Eastern Mediterranean. In case of predominantly subduction of oceanic crust, Mart (1994) expects an extensional regime for the northeastern African Plate.

First evidence for compression in the collisional zone of the NE African and Eurasian Plates seems to predate initial uplift in the Syrian Arc only slightly. The collisional zone is flanked by a characteristic Upper Cretaceous ophiolite belt extending from Cyprus to Oman. For Turkey, a compressional regime was postulated to have occurred after the Cenomanian or Turonian (Sengör & Yilmaz, 1981: 217; Collins & Robertson, 1997) while from the Oman mountains, Patton & O'Connor (1988) described the uplift of a swell during the Cenomanian and Turonian. Complete closure of the Birkat and Zagros suture lines is assumed to have taken place significantly later during the middle Eocene to late Miocene (Yilmaz, 1993; Searle, 1994: 1332; Yigitbas & Yilmaz, 1996). Some authors (Mart, 1994; Guiraud & Bosworth, 1996 and in press; Guiraud, in press; Guiraud & Bellion, 1995) assume that during the Late Cretaceous several short compressive phases were active in Northern Africa and Arabia and they reject the model of a long-lasting compressional regime. According to Guiraud and Bosworth (1996 and in press), Guiraud (in press) and Guiraud & Bellion (1995), one of the best documented compressional events occurred in the late Santonian for which they list evidence from Morocco, the Syrian Arc, Oman, the Benue/Tchad-intraplate-basin and other areas.
9.8.3. Campanian-Maastrichtian

At Areif El Naqa we reconstructed a late Campanian-early Maastrichtian uplift phase on the basis of siliciclastics which are interpreted to have been eroded from older siliciclastic deposits exhumated in the anticlinal core. It is unclear whether this uplift phase corresponds to a major folding phase postulated by Moustafa et al. (1991) and Moustafa & Khalil (1995) for Gebel Yelleq which is located approximately 80 km to the west of Gebel Areif El Naqa. These authors reconstructed that first major folding activities started during the late Senonian after deposition of the lower part of the Upper Senonian chalk which resulted in the formation of an intrachalk angular unconformity. Unfortunately the authors do not provide a biostratigraphic age data so that a direct correlation with our findings cannot be made.

Tectonic reconstructions from the Negev Desert of this period correspond well with the late Campanian-early Maastrichtian uplift phase at Areif El Naqa. An important tectonic peak for the middle Campanian is postulated in the Negev Desert by Zohar & Moshkovitz (1984), Honigstein et al. (1988) and Zur et al. (1995) on the basis of an unconformity at the Campanian-Maastrichtian boundary and biostratigraphically well constrained thickness variations. Besides the tectonic component, we expect the regional sea-level fall around the Campanian-Maastrichtian boundary (Ca/MaSin; Fig. 9-2; Lüning et al., submitted a) to have at least partly contributed to the formation of this unconformity.

The late Campanian-early Maastrichtian development in the Sinai-Negev Foldbelt contrasts the models of Guiraud & Bosworth (1996, and in press), Guiraud (in press) and Guiraud & Bellion (1995) who postulated a tethyan-wide rifting phase from the Campanian to the Maastrichtian or Paleocene and cite evidence from northern Libya, the southern Palmyrides and other regions. In Sinai the extensional phase is reported as having led to renewed subsidence during this time (Guiraud & Bosworth, in press and pers. comm. Guiraud, 1996) which does not match the situation at Areif El Naqa. For the late Maastrichtian in North Africa, Guiraud (in press) and Guiraud & Bellion (1995) describe a compressional phase which again does not match our observations at Areif El Naqa (see below).

9.8.4. Late Maastrichtian-Paleocene

At Areif El Naqa, the K/T boundary interval does not yield evidence for compressional movements during the late Maastrichtian and early Paleocene. However, a major tectonic uplift phase for the mid to late Paleocene is inferred on the basis of a biostratigraphically constrained unconformity at the northern anticlinal flank as discussed earlier in the text. Similar to the structural situation at the Areif El Naqa anticline, the Paleocene and lower Eocene rocks at the Gebel Yelleq anticline are flat-lying or show minor dips in contrast to the folded upper Cretaceous deposits. Moustafa et al. (1991) and Moustafa & Khalil (1995) interpreted this as an angular unconformity related to the termination of folding towards the end of the Cretaceous. Similar unconformities are known from the Mitla Pass in NW Sinai (Moustafa & Khalil, 1989) and from the area of Abu Roash, SW of Cairo (Moustafa, 1988; Hamza, 1993). At Areif El Naqa, however, we assume a continuous phase of tectonic activity in pulses during at least the Paleocene. It is clear that the oldest (Upper Cretaceous) strata was folded most intensely due to the syntectonic character of deposition. However, concerning the interpretation of the 'angular unconformity' it has to be considered that the bedding planes of the soft, homogenous Paleocene marls are often not clearly visible in the field. Furthermore the structure of the Areif El Naqa anticline with strongly tilted Cretaceous and only slightly dipping Paleocene strata may at least in part be interpreted as a pure structural phenomena connected with a strongly convergent folding style (class 1A of Ramsay, 1987). It is also important to note that the Paleogene strata in the domal anticlines of the Sinai Foldbelt are only preserved in the more distal parts of the anticlinal flanks so that the Paleogene tectono-sedimentary situation for the area near the anticlinal cores can only be approximated. The post-Cretaceous tectonic activity interpreted for the Areif El Naqa anticline is also supported by data from other locations in the Syrian Arc. From the Hatira monocline in the northern Negev Desert Zur et al. (1995) described an unconformity between the Paleocene Taqiyeh Fm. and the lower Eocene Mor Formation which corresponds well with the tectonically induced mid to upper Paleocene hiatus observed at Areif El Naqa. For the Gulf of Suez region, Patton et al. (1994; 21, additional references therein) described a major uplift phase which initiated in the upper Paleocene and continued through the early Eocene. A late Paleocene tectonic phase in Egypt was previously also postulated by Strougo (1986) which he termed 'velascoensis event'. Unfortunately he misinterpreted the tectonic movements as being associated with early rifting in the Red Sea-Gulf of Suez system which in the Gulf of Suez by most authors is assumed to have started not earlier than late Oligocene-early Miocene (e.g. Baldridge et al., 1991; Moustafa, 1993; Patton et al., 1994). Furthermore, at least part of Strougo's 'tectonic' evidence, which predominately is based on facies and thickness changes during the Gl. pseudomenardii and M. velascoensis Zones in the Western and Eastern Desert, may be explained by a low (eustatic) sea level during the Gl. pseudomenardii Zone leading to the development of lowstand massflows, the change from a formerly hemipelagic facies to laterally highly differentiated shallow marine subfacies, and amplified bottom currents resulting in non-deposition or even submarine erosion (see details in Lüning et al., submitted b).
For the Palmyride Foldbelt, a major period of shortening (15 km) has been postulated for the late Maastrichtian to late Eocene based on a distinct angular unconformity between the sedimentary complexes of the Cenomanian-Turonian and Maastrichtian-Eocene (Salel & Seguret, 1994). In contrast Chaimov et al. (1992) assume a period of tectonic quiescence for the Paleogene from seismic data, interrupted only in the middle Eocene by minor tectonism.

Summed up, the uplift phases reconstructed for Areif El Naqa correspond quite well with most tectonic reconstructions from other areas in the Syrian Arc. On the basis of the observations from Areif El Naqa combined with the listed literature data the main compressional phases for the Sinai-Negev Foldbelt are assumed for the Coniacian-early Santonian, late Campanian-early Maastrichtian and mid Paleocene-early Eocene. We expect that at least some of the discrepancies observed in the (inter-)regional correlation of the uplift phases may be related to misinterpretation of eustatic sea level drops, partly diachronous correlation of data from different biostratigraphic groups as well as from lithostratigraphic units, problems with differentiation between passive and actively uplifted relief, and to ambiguous continental-scale extrapolation of distinct tectonic phases including different complex orogenic systems.

9.9. Conclusions

The Areif El Naqa domal anticline in northeastern Sinai is part of the ‘Syrian Arc’ which represents an intraplate orogen that has formed since the late Cretaceous by inversion of an older halfgaben system as a consequence of the collision of the African and Eurasian Plates. The depositional history at Areif El Naqa has been reconstructed in terms of sequence stratigraphy on the basis of detailed sedimentologic, biostratigraphic and paleoecologic investigations in ten sections as well as literature data of the pre- and syndeformational upper Albian to lower Eocene succession. The late Albian-late Turonian sediments were deposited under technically quiet conditions and were influenced mainly by interregional sea-level variations. For the Senonian to early Eocene three significant uplift phases have been reconstructed for the Areif El Naqa anticline based on evidence from lateral facies and thickness changes, local development of pronounced hiatuses and comparison with the sequence stratigraphic development in the tectonically quiet region of central east Sinai. The first tectonic phase is interpreted for the Coniacian-early Santonian and is characterized by pronounced facies and thickness changes which were documented in an earlier study by Bartov et al. (1980). Nevertheless, regional sea level changes still played a significant role for deposition at Areif El Naqa during this period. The second tectonic phase is indicated for the late Campanian-early Maastrichtian by siliciclastics which are interpreted to have been reworked from older siliciclastics uplifted in the anticlinal core. The third compressional period is assumed for the middle Paleocene to early Eocene as evidenced by a significant hiatus in sections at the anticlinal flank. The uplift history at Areif El Naqa has been compared with the tectonic development in other parts of the Syrian Arc and in general seems to reflect major movements which occurred throughout the anticlines of the foldbelt.
References


Hancock, J. M. (1993): Comments on the Exxon cycle chart for the Cretaceous system. Cuad. de Geol. Iberica 17, 57-78.}


Appendix A

Compilation of references for the Cenomanian to Eocene of Sinai
-sorted by area-

(contributions dealing with Sinai as a whole are not listed
but are cited in the earlier chapters of this thesis)
Regional distribution of published studies from Sinai and the Negev dealing with the period Cenomanian to Eocene. Numbers refer to tables on the following pages. Numbers in brackets indicate study areas that are located in more than one stippled field.
Bibliography for the Cenomanian-Eocene of Sinai. References are arranged by number as used in the map on the previous page. Listed are all publications dealing with the regional geology of Sinai for the interval Cenomanian to Eocene (as far as known and accessible). G. = Gebel (Mountain), W. = Wadi.

<table>
<thead>
<tr>
<th>No.</th>
<th>Reference</th>
<th>Area of study</th>
<th>Studied interval</th>
<th>Subject</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Abdel Gawad (1990)</td>
<td>Safar, W. Qasr Mina, Tabia</td>
<td>Cenomanian to Santonian</td>
<td>stratigraphy, biostratigraphy (plankton)</td>
</tr>
<tr>
<td>2</td>
<td>Abdel Gawad &amp; Zalat (1992)</td>
<td>G. El-Hamra, G. Heriba (Mitla-Pass)</td>
<td>Cenomanian to Turonian</td>
<td>stratigraphy, biostratigraphy (plankton)</td>
</tr>
<tr>
<td>3</td>
<td>Abdel-Gawad et al. (1992)</td>
<td>G. Nezzarat</td>
<td>Cenomanian to Turonian</td>
<td>biostratigraphy (plankton, forams.)</td>
</tr>
<tr>
<td>4</td>
<td>Abdelmalik et al. (1978a+b)</td>
<td>Bir El-Marka</td>
<td>Maastrichtian to Eocene</td>
<td>biostratigraphy (plankton, forams., nannofossils)</td>
</tr>
<tr>
<td>5</td>
<td>Abdou &amp; Fawzi (1966)</td>
<td>G. El Minshoer</td>
<td>Cenomanian</td>
<td>biostratigraphy (macroinvertebrates)</td>
</tr>
<tr>
<td>6</td>
<td>Allam &amp; Khalil (1988)</td>
<td>G. Areif El Naga</td>
<td>Triassic to Eocene</td>
<td>stratigraphy, tectonics (review article, few new results)</td>
</tr>
<tr>
<td>7</td>
<td>Allam &amp; Khalil (1989)</td>
<td>G. Qabeliat</td>
<td>Cambrian to recent</td>
<td>mapping, stratigraphy, facies development.</td>
</tr>
<tr>
<td>11</td>
<td>Bartov et al. (1972)</td>
<td>Sudl. Negev (incl. Tabia)</td>
<td>Cenomanian to Eocene</td>
<td>stratigraphy, facies development, updated in Bartov &amp; Sternitz (1977)</td>
</tr>
<tr>
<td>12</td>
<td>Bartov et al. (1980)</td>
<td>G. Areif El Naga</td>
<td>Triassic to Noogene</td>
<td>stratigraphy, facies development, tectonics</td>
</tr>
<tr>
<td>13</td>
<td>Braun et al. (1987)</td>
<td>northern Negev</td>
<td>Turonian</td>
<td>stratigraphy, facies development, tectonics</td>
</tr>
<tr>
<td>15</td>
<td>Cherif et al. (1989a)</td>
<td>G. Nezzarat, G. Makkat, G. Ekma, G. Qabiliat</td>
<td>Cenomanian to Turonian</td>
<td>biostratigraphy (forams.), palaeoecology</td>
</tr>
<tr>
<td>16</td>
<td>Cherif et al. (1989b)</td>
<td>SE’ Abu Rudeis</td>
<td>Campanian to Maastrichtian</td>
<td>biostratigraphy (plankton, forams.)</td>
</tr>
<tr>
<td>17</td>
<td>Colin &amp; El-Dakkak (1975)</td>
<td>G. Nezzarat</td>
<td>Campanian</td>
<td>systematics ostracodes</td>
</tr>
<tr>
<td>18</td>
<td>El Aassy (1992)</td>
<td>E’ Qaa-Ebene</td>
<td>Campanian to Maastrichtian</td>
<td>stratigraphy, petrology (uranium in phosphorites)</td>
</tr>
<tr>
<td>20</td>
<td>El Beify (1993)</td>
<td>Wadi el-Esela</td>
<td>Offshore Nordsina</td>
<td>Alban to Cenomanian</td>
</tr>
<tr>
<td>23</td>
<td>El Taraibbi &amp; Adawy (1972)</td>
<td>S’ Abu Zenima</td>
<td>Phaseozoic</td>
<td>tectonics</td>
</tr>
<tr>
<td>25</td>
<td>Givitzman et al. (1989)</td>
<td>Central Israel</td>
<td>Turonian to Maastrichtian</td>
<td>biostratigraphy, stratigraphy</td>
</tr>
<tr>
<td>26</td>
<td>Kassab &amp; Ismael (1994)</td>
<td>G. Musaba Salam</td>
<td>Cenomanian to Santonian</td>
<td>systematics (macroinvertebrates), palaeobathmetry</td>
</tr>
<tr>
<td>27</td>
<td>Haggag &amp; Lutterbacher (1991)</td>
<td>W. Nukhul</td>
<td>Eocene</td>
<td>biostratigraphy (plankton, forams.)</td>
</tr>
<tr>
<td>28</td>
<td>Hassan et al. (1992)</td>
<td>G. Risan Aneiza, G. El Amrur</td>
<td>Aptian to Turonian</td>
<td>carbonate microfossils, geochemistry</td>
</tr>
<tr>
<td>29</td>
<td>Hawiady (1987)</td>
<td>El Qusaima</td>
<td>Palaeocene to Eocene</td>
<td>biostratigraphy, paleoecology</td>
</tr>
<tr>
<td>30</td>
<td>Hawiady &amp; El Ashaw (1993)</td>
<td>El Qusaima</td>
<td>Turonian to Maastrichtian</td>
<td>biostratigraphy (plankt. + benth. forams.), palaeobathmetry</td>
</tr>
<tr>
<td>31</td>
<td>Hawiady et al. (1991)</td>
<td>El Qusaima</td>
<td>Turonian to Maastrichtian</td>
<td>biostratigraphy (plankt. + benth. forams.)</td>
</tr>
<tr>
<td>32</td>
<td>Honigstein et al. (1988)</td>
<td>northern Negev</td>
<td>Senonian</td>
<td>tectonics, biostratigraphy (ostracodes)</td>
</tr>
<tr>
<td>33</td>
<td>Imnael (1992)</td>
<td>G. Musaba Salam, G. Marifat, G. Qabiliat</td>
<td>Maastrichtian to Eocene</td>
<td>biostratigraphy (benth. forams.), palaeoecology</td>
</tr>
<tr>
<td>34</td>
<td>Jenkins et al. (1982)</td>
<td>G. Magbara</td>
<td>Jurassic to Eocene</td>
<td>mapping et.</td>
</tr>
<tr>
<td>35</td>
<td>Keller &amp; Benjamins (1991)</td>
<td>Ngev: Zofar, Hor Hahar, El Morr, Ben Gurion, near El Qusaima</td>
<td>Jurassic to Eocene</td>
<td>biostratigraphy (plankt. forams.), sea level changes</td>
</tr>
<tr>
<td>36</td>
<td>Kora &amp; Ayyad (1988)</td>
<td>Bir Safra</td>
<td>Eocene</td>
<td>biostratigraphy, biofacies, tectonics</td>
</tr>
<tr>
<td>38</td>
<td>Kora et al. (1994)</td>
<td>W. Abu Qada, G. Musaba Salam, G. Farah El Ghozlan, W. Feiran</td>
<td>Cenomanian</td>
<td>biostratigraphy (forams.), palaeoecology</td>
</tr>
<tr>
<td>39</td>
<td>Marzouk &amp; Abou-El-Enein (1995)</td>
<td>W. Feiran, G. Qabiliat</td>
<td>Maastrichtian to Eocene</td>
<td>biostratigraphy (nannofossils)</td>
</tr>
<tr>
<td>40</td>
<td>Moray et al. (1995)</td>
<td>G. Dushal, G. Shiti, G. Gunna</td>
<td>Triassic to Lower Cretaceous</td>
<td>sedimentology of siliciclastics</td>
</tr>
<tr>
<td>41</td>
<td>Moustafa &amp; Khalil (1994)</td>
<td>Wadi El-Esseiia</td>
<td>Triassic to Eocene</td>
<td>sedimentology of siliciclastics</td>
</tr>
<tr>
<td>42</td>
<td>Moustafa &amp; Khalil (1994)</td>
<td>Themed Fault</td>
<td>Lower Mesozoic to recent</td>
<td>tectonics</td>
</tr>
<tr>
<td>No.</td>
<td>Reference</td>
<td>Area of study</td>
<td>Studied interval</td>
<td>Subject</td>
</tr>
<tr>
<td>-----</td>
<td>-----------</td>
<td>---------------</td>
<td>------------------</td>
<td>---------</td>
</tr>
<tr>
<td>44</td>
<td>Moustafa &amp; Khalil (1995)</td>
<td>G. Yeleq</td>
<td>Upper Cretaceous to Miocene</td>
<td>tectonics</td>
</tr>
<tr>
<td>45</td>
<td>Moustafa et al. (1991)</td>
<td>G. Yeleq</td>
<td>Upper Cretaceous to Lower Tertiary</td>
<td>tectonics</td>
</tr>
<tr>
<td>46</td>
<td>Nakkady (1950)</td>
<td>various sections, incl- Sinai: W. Danili, Abu Durbah</td>
<td>Senonian to Eocene</td>
<td>systematics, biostratigraphy (plankt. + benth. forams.)</td>
</tr>
<tr>
<td>47</td>
<td>Omara (1956)</td>
<td>G. Nezzazat</td>
<td>Cenomanian</td>
<td>systematics (forams.)</td>
</tr>
<tr>
<td>48</td>
<td>Orabi (1992)</td>
<td>W. Wair</td>
<td>Cenomanian to Turonian</td>
<td>biostratigraphy (forams.)</td>
</tr>
<tr>
<td>49</td>
<td>Parnes (1987)</td>
<td>G. E Ze-Roha</td>
<td>Upper Turonian</td>
<td>systematics (rudists)</td>
</tr>
<tr>
<td>50</td>
<td>Philipps (1994)</td>
<td>10 km SW Thamed, 20 km E' Abu Zenima</td>
<td>Maastrichtian to Eocene</td>
<td>biostratigraphy (coccol.), geochemistry</td>
</tr>
<tr>
<td>51</td>
<td>Sait &amp; Kessawy (1956)</td>
<td>Nakhl, Giddi</td>
<td>Maastrichtian to Paleocene</td>
<td>systematics + biostratigraphy (forams.), basin development</td>
</tr>
<tr>
<td>52</td>
<td>Shahin (1990)</td>
<td>G. Ekma</td>
<td>Maastrichtian to Eocene</td>
<td>benth. forams., paleochemistry</td>
</tr>
<tr>
<td>53</td>
<td>Shahin (1991)</td>
<td>G. Nezzazat</td>
<td>Cenomanian to Turonian</td>
<td>systematics (ostracodes), geochemistry</td>
</tr>
<tr>
<td>55</td>
<td>Shahin &amp; Kora (1991)</td>
<td>Central Eastern Sinai, various localities</td>
<td>Cenomanian to Maastrichtian</td>
<td>biostratigraphy (plankt. + benth. forams.), paleochemistry</td>
</tr>
<tr>
<td>56</td>
<td>Soudry (1987)</td>
<td>northern Negev</td>
<td>Campanian</td>
<td>petrology, phosphorites</td>
</tr>
<tr>
<td>57</td>
<td>Strougo &amp; Hamza (1989)</td>
<td>Bir Haleifya</td>
<td>Eocene</td>
<td>systematics (macroinvertebrates), biostratigraphy, stratigraphy</td>
</tr>
<tr>
<td>58</td>
<td>Yousef &amp; Shinnawi (1954)</td>
<td>W. Sudr</td>
<td>Turonian to Eocene</td>
<td>stratigraphy, tectonics</td>
</tr>
<tr>
<td>59</td>
<td>Ziko et al. (1993)</td>
<td>Central Eastern Sinai, various localities</td>
<td>Cenomanian to Eocene</td>
<td>stratigraphy, little facies development.</td>
</tr>
<tr>
<td>60</td>
<td>Sayed (1990)</td>
<td>G. Nezzazat</td>
<td>Cenomanian to Turonian (?)</td>
<td>biostratigraphy</td>
</tr>
<tr>
<td>61</td>
<td>Hamza et al. (1994)</td>
<td>G. Manzour, G. El Minskrah</td>
<td>Aptian to Turonian</td>
<td>biostratigraphy, paleoecology</td>
</tr>
<tr>
<td>62</td>
<td>Steinitz (1970)</td>
<td>Southern Negev (near Ellat)</td>
<td>Senonian</td>
<td>sedimentary structures in chert</td>
</tr>
<tr>
<td>63</td>
<td>Ahmed (1995)</td>
<td>Taba</td>
<td>Cenomanian to Maastrichtian</td>
<td>microfacies</td>
</tr>
<tr>
<td>64</td>
<td>El Shiekh (1995)</td>
<td>Taba</td>
<td>Cenomanian to Maastrichtian</td>
<td>biostratigraphy (forams. + nannofossils)</td>
</tr>
<tr>
<td>65</td>
<td>Ismail (1993)</td>
<td>1 km S' Old Abu Zenima</td>
<td>Coniacian to Santonian</td>
<td>biostratigraphy + systematics (forams. + ostracodes)</td>
</tr>
<tr>
<td>66</td>
<td>Anan (1992)</td>
<td>W. Nukhul</td>
<td>Maastrichtian to Lower Eocene</td>
<td>biostratigraphy (plankton, forams.)</td>
</tr>
<tr>
<td>67</td>
<td>Abdel Gawad &amp; Gameld (1992)</td>
<td>G. Nezzazat</td>
<td>Cenomanian</td>
<td>systematics (gastropodes)</td>
</tr>
<tr>
<td>70</td>
<td>Hawwaiby (1993)</td>
<td>offshore NW-Sinai</td>
<td>Aptian to Turonian</td>
<td>biostratigraphy (plankt. forams.), Paleobathymetry</td>
</tr>
<tr>
<td>71</td>
<td>El Kelany &amp; Said (1986-89)</td>
<td>SE-Sinai</td>
<td>Precambrian to Recent</td>
<td>lithostratigraphy</td>
</tr>
<tr>
<td>72</td>
<td>Orabi (1995)</td>
<td>W. Wair, W. Taba</td>
<td>Cenomanian to Lower Turonian</td>
<td>biostratigraphy (Macroinvertebrates), paleoecology</td>
</tr>
<tr>
<td>75</td>
<td>Orabi (1992)</td>
<td>Drilling in Abu Rudeis-Belayim-area</td>
<td>Cenomanian-Maastrichtian</td>
<td>biostratigraphy (plankt.+benth. forams.), paleochemistry</td>
</tr>
<tr>
<td>76</td>
<td>Allam et al. (1986)</td>
<td>G. Qabeliat</td>
<td>Campanian to Maastrichtian</td>
<td>biostratigraphy (plankt. forams., Nanofosses.)</td>
</tr>
<tr>
<td>77</td>
<td>Faris et al. (1986)</td>
<td>G. Qabeliat</td>
<td>Eocene</td>
<td>biostratigraphy (plankt. forams., Nanofosses.)</td>
</tr>
<tr>
<td>78</td>
<td>El-Shinnawi (1967)</td>
<td>W. Sudr</td>
<td>Maastrichtian</td>
<td>biostratigraphy (plankton. forams., microfacies)</td>
</tr>
<tr>
<td>79</td>
<td>El-Walid (1994)</td>
<td>Bellayim + Rudeis oilfield</td>
<td>Coniacian</td>
<td>facies model</td>
</tr>
<tr>
<td>80</td>
<td>Orabi &amp; Ramadan (1995)</td>
<td>W. Umm, W. Abuira</td>
<td>Coniacian</td>
<td>phosphorites</td>
</tr>
<tr>
<td>81</td>
<td>Soudry (1992)</td>
<td>Central Negev</td>
<td>Senonian</td>
<td>phosphatic omission surface</td>
</tr>
<tr>
<td>82</td>
<td>Soudry &amp; Lewy (1990)</td>
<td>Nahal Zinim (Southern Negev)</td>
<td>Campanian</td>
<td>phosphatic omission surface</td>
</tr>
<tr>
<td>83</td>
<td>Ayad et al. (1996)</td>
<td>G. Areif El Naqa</td>
<td>Barremian to Maastricht.</td>
<td>biostratigraphy (plankton. forams.)</td>
</tr>
<tr>
<td>84</td>
<td>Halkina &amp; Eid (1995)</td>
<td>Central Eastern Sinai</td>
<td>Coniacian to Campanian</td>
<td>facies, sea level, geodynamics</td>
</tr>
<tr>
<td>85</td>
<td>Abbass et al. (1994)</td>
<td>West Central Sinai</td>
<td>Maastr. to Lower Eocene</td>
<td>benthon. forams., biostrat., paleoecology</td>
</tr>
</tbody>
</table>
Appendix B

Summarizing diagram of the biozonal schemes of planktonic foraminifera and calcareous nannofossils as used in this study
**Biozonal schemes of planktonic foraminifera and calcareous nannofossils as used in the study and their correlation**

Boundary ages and their correlation with foraminiferal biozones after Gradstein et al. (1995) and Boill et al. (1985) (Cretaceous), and Berggren et al. (1985) (Paleogene). Nannofossil zones are directly correlated with foraminiferal zones as based on the studied sections from Eastern Sinai and literature data. Biozones which were not recorded are shaded.
Appendix C

Accumulation rates during the late Maastrichtian and Paleocene
Average accumulation rates

<table>
<thead>
<tr>
<th>Biozone</th>
<th>Duration / my</th>
<th>Section A7</th>
<th>Section A8</th>
<th>Section C</th>
<th>Section F</th>
<th>Section P</th>
<th>Section Q</th>
<th>Section T1</th>
</tr>
</thead>
<tbody>
<tr>
<td>G. pseudomenardii</td>
<td>3.3</td>
<td>NX</td>
<td>IC</td>
<td>10.5</td>
<td>5.3</td>
<td>5.3</td>
<td>5.3</td>
<td>5.3</td>
</tr>
<tr>
<td>L. pusilla pusilla</td>
<td>1.0</td>
<td>NX</td>
<td>3.0</td>
<td>9.6</td>
<td>4.1</td>
<td>4.1</td>
<td>4.1</td>
<td>4.1</td>
</tr>
<tr>
<td>M. angulata</td>
<td>0.7</td>
<td>NX</td>
<td>2.3</td>
<td>9.2</td>
<td>2.2</td>
<td>2.2</td>
<td>2.2</td>
<td>2.2</td>
</tr>
<tr>
<td>P. uncinata</td>
<td>0.5</td>
<td>NX</td>
<td>10.5</td>
<td>8.7</td>
<td>13.6</td>
<td>27.2</td>
<td>13.6</td>
<td>27.2</td>
</tr>
<tr>
<td>P. trinidadensis</td>
<td>1.1</td>
<td>NX</td>
<td>8.6</td>
<td>7.8</td>
<td>7.8</td>
<td>7.8</td>
<td>7.8</td>
<td>7.8</td>
</tr>
<tr>
<td>P. pseudobulloides</td>
<td>2.4</td>
<td>NX</td>
<td>IC</td>
<td>3.0</td>
<td>1.3</td>
<td>&gt;3.7</td>
<td>&gt;1.5</td>
<td>&gt;5.0</td>
</tr>
<tr>
<td>A. mayaroensis</td>
<td>2.2</td>
<td>29.4</td>
<td>13.4</td>
<td>IC</td>
<td>IC</td>
<td>&gt;25.9</td>
<td>&gt;11.8</td>
<td>&gt;28.3</td>
</tr>
</tbody>
</table>

Table: Thicknesses and average accumulation rates (for non-decompacted sediments) for different planktonic foraminiferal biozones of selected sections.

The average accumulation rates have been calculated for different planktonic foraminiferal biozones of selected sections (location map in Fig. 1-1). The values refer to non-decompacted sediments. The highest values have been noticed for the P. uncinata Zone, the lowest for the P. pseudobulloides Zone. Clear relationships between the average accumulation rates and environmental parameters, such as changes in sea level changes or paleoproductivity, are not developed or cannot be resolved. The values lie within the typical range for (hemi-) pelagic accumulation rates. For example, Tucker & Wright (1990) report sedimentation rates for non-compacted pelagic carbonates of the order of 10-50 mm/1000 years. Similar values are listed in Enos (1991) for modern pelagic carbonates and have been also reconstructed for the Late Cretaceous Chalk in the UK (3-60 mm 1000 yrs) and in the Western Interior, USA (6.5-50 mm 1000 yrs) (Scholle et al., 1983).

Interpretation of the dataset is complicated by a relatively high number of uncertainties and processes:

- Sample spacing and therefore resolution lies between 1-2 m.
- The exact tilt (and therefore thickness) of subhorizontal strata is sometimes complicated to determine in marly outcrops because of the homogenous lithology and a thin cover of erosional debris.
- Within the A. mayaroensis, M. angulata, G. pseudomenardii and M. velascoensis Zones hiatuses or firmgrounds/hardgrounds may be locally or regionally developed (see chapter 7). The calculated accumulation rates are averaged over composite periods consisting of deposition, non-deposition and even erosion. Often, the different phases can biostratigraphically not be resolved (see also e.g. Moore et al., 1978).
- The view of hemipelagic and pelagic sedimentation as a slow, steady rain of debris is obsolete and inaccurate as sediment trap records have shown (see Grimm et al. 1996 and references therein). In the study area this observation is especially important for the Late Maastrichtian where short-termed low and high productivity phases alternated (see chapter 6). Accumulation rates are expected to have been significantly higher during the high productivity phases. The calculated accumulation rates for the A. mayaroensis Zone give only averaged values without differentiation between high and low productivity conditions. In sections F and P the base of the A. mayaroensis Zone is not exposed so that only minimum values can be given.
- Bottom currents and the paleorelief may contribute to lateral changes in the accumulation rates.
- Major uncertainties are expected in the chronostrati-graphic dating of microfossil events (see also chapter 8) and their global validity. It need not to be said that the calculated rates depend strongly on the correct chronological ages of the different biostratigraphic zones. This also seems to play a role for the rates calculated, for example the P. uncinata Zone is often marked by suspicious high values.
Appendix D

Sections measured in eastern Sinai

(legend on next page, location map in Fig. 1-1 / chapter 1.3.)
Legend for sections
dolomitic limestone with chert layers/nodules

100%

Foraminifera

100%

Section A1b: Gebel Areif El Naqa
30°23'11" N, 34°28'21.9" E
Section A3a: Gebel Areif El Naqa
Turanian (/ Coniacian)
Wata / Themed Fm. (Geroft / Zihor Fm.)

(no biostratigraphic data)
hard dolomitic chalk

hard thick bedded chalks
white, no marly interlayers

marly interlayer: high P/B-ratio

thick bedded chalks with thin marly interlayers

Section A4: Gebel Areif El Naqa
Section A5: Gebel Areif El Naqa
30°23'31.7" N, 34°29'10.6" E
G. falsostuarti  
G. gansseri  
Abathomphalus mayaroensis

A. cymbiformis  
L. quadratus

Micula murus

Foraminifera with flint layers
massive dolomite
without flint
alternated bedding of dolomite platy dolomitic mart
alternated bedding of dolomitic micrite and marl

Quart grains
In washing residue
(fraction 25(Kx<630)-A7-23 few %
- A7-22 Qz
- A7-21 A7-20 A7-19 A7-18 A7-17 A7-16 A7-15 A7-14 A7-13 A7-12 A7-11A7-10 A7-9

HST
TST

Section A7: Gebel Areif El Naga

30°24'31,1" N, 34°29,34,6" E
Section A8: Gebel Areif El Naqa

located a few 100 m near section A7
Section B (lower part): 20 km N' Sheikh Attiya

Section B' (approx. 500 m from section B): 20 km N' Sheikh Attiya
Section B (middle part): 20 km N' Sheikh Attiya
Section B (uppermost part): 20 km N' Sheikh Atiya
Section C: 5 km N' Sheikh Attiya
29°16'13.1"N / 34°30'01.9"E
Section D: 20 km N' Sheikh Attiya

Section F: Gebel Misheiti, 10 km NNE' Themed
Section G: 25 km SE' Themed

Section H: 30 km SE' Themed
Section M (uppermost part): Taba

Enlargement of Coniacian siliciclastics in the Taba section
(meters at scale bar refer to the main section illustrated before)
Enlargement of the Campanian/Maastrichtian boundary in the Taba section
(meters at scale bar refer to main section illustrated before)

Enlargement of the lithological change from siliciclastics to limestones/cherts in the Coniacian/Santonian of the Taba section
(meters at scale bar refer to main section illustrated before)
### Section P: Wadi Gureis

**Map and Data:**
- **Location:** 29°33'41.7" N / 34°09'08.0" E

**Rock Units:**
- Dolomitic limestone with thin interbeds
- Soft calcareous marl (P1-35) over Dioturbate hard calcareous marl over green shale (P1-34 and P1-33) over hard dolomitic calcareous marl.

**Fauna:**
- **Foraminifera:**
  - Foraminifera planktonic (100%)
  - Foraminifera benthonic (100%)

**Fossil Assemblages:**
- M. uncinata
- M. angularis
- M. pseudomenardii
- P. uncinata
- P. tripladralis

**Section:**
- **P1-2 to P1-15**
  - Lower Maastrichtian
  - Upper Maastrichtian

**Geological Units:**
- **Esna Shale**
  - Jurassic
- **Dolomitic limestone**

**Age Determination:**
- **Planktonic Foraminifera Zones:**
  - NP1 to NP1-35

### Section Q: Egma Plateau

**Map and Data:**
- **Location:** 29°21'54.8" N / 34°04'56.5" E

**Rock Units:**
- Dolomitic limestone, base with thinbedded calcareous marl
- Greenish marl and calcareous marl
- Hard calcareous marl
- Greenish marl and calcareous marl

**Fauna:**
- **Foraminifera:**
  - Foraminifera planktonic (100%)
  - Foraminifera benthonic (100%)

**Fossil Assemblages:**
- M. subbotinii
- M. pseudomenardii
- P. uncinata
- P. trilobata

**Section:**
- **Q1-1 to Q1-35**
  - Lower Cretaceous
  - Upper Cretaceous

**Geological Units:**
- **Esna Shale**
- **Dolomitic limestone**

**Age Determination:**
- **Planktonic Foraminifera Zones:**
  - NP1 to NP1-35
**Section R: N’ Gebel Umm Mafrud/Egma Plateau**

29°10'44,0" N / 34°0,9" E

**Section T1: between Nuweiba and Taba**

29°16'43,4" N / 34°43'47,9" E
Section W: between G. Areif El Naga and El Kuntilla
30°08'23.2" N / 34°35'49.0" E