The Levant Basin Offshore Israel: Stratigraphy, Structure, Tectonic Evolution and Implications for Hydrocarbon Exploration

Michael Gardosh, Yehezkel Druckman, Binyamin Buchbinder and Michael Rybakov

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The Levant Basin Offshore Israel: Stratigraphy, Structure, Tectonic Evolution and Implications for Hydrocarbon Exploration

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Abstract

The Levant Basin is a deep and long existing geologic structure located in the eastern Mediterranean Sea. The southern part of the basin hosts a world-class hydrocarbon province offshore the Nile Delta. Recent discoveries of biogenic gas and various oil shows indicate that the central part of the basin, offshore Israel has significant hydrocarbon potential. To further investigate this area, which is generally under-explored, a basin-scale study was initiated by the petroleum commissioner in the Israeli Ministry of Infrastructure. In this study new geophysical data is integrated with regional and local geologic data to reconstruct the basin history and to identify favorable hydrocarbon plays.

The reconstructed basin history shows that the Levant Basin was shaped in several main tectonic stages. Early Mesozoic rifting resulted in the formation of an extensive graben and horst system, extending throughout the Levant onshore and offshore. Late Jurassic to Middle Cretaceous, post-rift subsidence was followed by the formation of a deep marine basin in the present-day offshore and a shallow-marine, carbonate dominated margin and shelf near the Mediterranean coastline and further inland. A Late Cretaceous and Tertiary convergence phase resulted in inversion of Early Mesozoic structures and the formation of extensive, Syrian Arc-type contractional structures throughout the Levant Basin and margin. The Tertiary convergence was further associated with uplift, widespread erosion, slope incision and basinward sediment transport. A Late Tertiary desiccation of the Mediterranean Sea was followed by deposition of a thick evaporitic blanket that was later covered by a Plio-Pleistocene siliciclastic wedge. The Phanerozoic basin-fill ranges in thickness from 5-6 km on the margin to more than 15 km in the central part of the basin.

A variety of potential, structural and stratigraphic traps were formed in the Levant Basin during the three main tectonic stages. Possible hydrocarbon play types are: Triassic and Lower Jurassic fault-controlled highs and rift-related traps; Middle Jurassic shallow-marine reservoirs; Lower Cretaceous deepwater siliciclastics; the Tertiary canyon and channel system and associated deepwater siliciclastics found in either confined or non-confined settings; Upper Cretaceous and Tertiary Syrian Arc folds; and Mesozoic and Cenozoic isolated, carbonate buildups on structurally elevated blocks.

Shallow gas discoveries in Pliocene sands and high-grade oil shows found in the Mesozoic section indicate the presence of source rocks and appropriate conditions for hydrocarbon generation in both biogenic and thermogenic petroleum systems. The size, depth and trapping potential of the Levant Basin suggest that large quantities of hydrocarbons can be found offshore Israel.
1. Introduction

The Levant Basin occupies the eastern Mediterranean Sea (Fig. 1.1). Its southern part, offshore Egypt hosts the prolific Nile Delta hydrocarbon province where extensive exploration and production activity is taking place. The increase in demand and rising price of oil in recent years has promoted a growing interest in the less known and under-explored, central and northern parts of the basin, offshore Israel, Lebanon and Cyprus.

Petroleum exploration in the eastern Mediterranean has a long history and has so far yielded modest success. Offshore exploration started in the late 1960's and early 1970's with a series of wells drilled by Belpetco (Fig. 1.2). The seven Belpetco wells targeted structural culminations on the shallow shelf of Israel and northern Sinai, but all were found dry. These early wells, however, provided important information and established the initial geologic model of the eastern Mediterranean as a deep marine basin that was formed during the Early Mesozoic, and continued to develop west of an extensive shallow-marine shelf.

The next exploration campaign during the mid 1970's to mid 1980's resulted in more success. Several wells were drilled on shallow structures offshore Sinai (Fig. 1.2). Light oil was found in Early Cretaceous sandstone in the Ziv-1 (drilled by Oil Exploration Limited) and the Mango-1 (drilled by Total) wells. Mango-1 initially produced 41-50\(^0\) API gravity oil at a rate of about 10,000 Bbl/day; however, performance rapidly decreased and production was stopped.

A series of four wells were drilled offshore Israel by Isramco during the late 1980's to late 1990's (Fig. 1.2). These boreholes that targeted structural culminations were typically deeper than earlier wells and most of them reached a Middle Jurassic stratigraphic level at depths of more than 5000 m. All the Isramco wells encountered various oil and gas shows. Yam-2 and Yam Yafo-1 tested 44-47\(^0\) API gravity oil at a rate of 500-800 Bbl/day from Middle Jurassic limestone, although no commercial production was established.

Exploration activity in the Levant Basin underwent a significant resurgence since 1999-2000 when several, large gas fields were discovered at a shallow depth within Pliocene sands west of the towns of Ashqelon and Gaza (Fig. 1.2). Gas reserves in the Noa/MariB/Gaza Marine fields are estimated at about 3 TCF. Gas is presently produced from the MariB field and is transported onshore to various Israeli consumers.

The successful exploration campaign of the early 2000's promoted the acquisition of geophysical data throughout the entire eastern Mediterranean area. The new geophysical data sets include about 12,000 km of regional 2D seismic reflection lines (Fig. 1.3) and several, high resolution 2D and 3D surveys, as well as new gravity and magnetic measurements.

The petroleum commissioner in the Israeli Ministry of Infrastructure initiated a regional analysis of the Levant Basin and margin. The study was carried out by professional teams from
Fig. 1.1- Main physiographic and tectonic elements of the Levant region. A-B indicates the location of the schematic cross-section in Fig. 2.1.
Fig. 1.2- Offshore drilling history and main hydrocarbon occurrences. The Noa-Or, Marie, Nir and Gaza Marine fields are recent gas discoveries in Pliocene deepwater sands.
Fig 1.3 Location of seismic lines, wells, and stratigraphic cross-sections (see Figs 4.2 and 4.6).

The map coordinates are in WGS 84, UTM Zone 31 projection.
the Geophysical Institute of Israel and Geological Survey of Israel. Its main emphasis was to integrate the vast amount of geological information from the Levant margin onshore with the high-quality geophysical data sets that were recently acquired in the basin offshore. The final results include depth maps of key stratigraphic horizons, stratigraphic cross-sections, and regional tectonic and paleogeographic maps. The results enabled the reconstruction of a regional geologic framework for the two areas, traditionally studied in different scales and techniques.

2. Tectonic Setting

Wide-angle deep refraction profiles show significant variations in seismic velocities and depth to the Moho from the inland part of Israel on the east to the Mediterranean Sea on the west (Makris et al., 1983; Ginzburg and Ben-Avraham, 1987, Ben-Avraham et al., 2002). The velocity variations (summarized in Fig. 2.1) are associated with a change from a ~30 km thick, low velocity continental-type crust inland, to a ~10 km thick oceanic- or intermediate-type crust offshore (Garfunkel, 1998, Ben-Avraham et al, 2002; Gardosh and Druckman, 2006). This fundamental geophysical observation and its interpretation are milestones in the research of the Levant area. They unequivocally demonstrate that the present-day marine basin is a deeply rooted geologic structure associated with large-scale plate motions.

The Paleozoic section found in outcrops and wells in southern and central Israel comprises continental to shallow marine deposits. Evidence from adjacent countries further indicates that during most of the Paleozoic the Levant region formed part of the super-continent Gondwana and was located inland, several hundred kilometers south of the Paleotethys Ocean (Garfunkel, 1998, Weissbrod, 2005). The Paleozoic landscape was dominated by a series of region-wide swells and intervening basins that were hundreds of kilometers in diameter (Weissbrod, 2005). The area of the eastern Mediterranean formed part of this landscape and no deep marine basin was in existence at that time.

The Late Permian, Triassic and Early Jurassic section of the Levant region is dominated by shallow-marine carbonate and occasionally siliciclastic rock. Significant thickness variations in the Triassic and Early Jurassic strata are identified in well and seismic data from southern and central Israel (Fig. 2.1) (Goldberg and Friedman, 1974; Druckman, 1974, 1977; Freund et al., 1975; Druckman, 1984; Bruner, 1991; Druckman et al., 1995a; Garfunkel, 1998, Gelberman, 1995). The Palmyra trough in central Syria (Fig. 1.1) contains an unusually thick Permian to Triassic section. This range of phenomena indicates extensive vertical motions, continental
Fig. 2.1 - Schematic section across the Levant Basin and margin, from the Eratosthenes Seamount to the Dead Sea Rift. The section shows the main structural elements, stratigraphic units and crustal seismic velocities (after Garfunkel, 1998). See Fig. 1.1 for location.
breakup and rifting of the Gondwanian shallow-shelf (Freund et al., 1975; Garfunkel and Derin, 1984, Garfunkel, 1998).

A series of horst and graben structures (Fig. 2.1) evolved in several pulses during the uppermost Paleozoic and Early Mesozoic periods, accompanied by widespread magmatic activity. These regional extensional pulses resulted in separation of the Tauride and Eratosthenes continental blocks from the Gondwanaland craton and the formation of a deep marine basin in the area of the present-day eastern Mediterranean (Bein and Gvirtzman, 1977; Garfunkel and Derin, 1984; Garfunkel, 1998; Robertson, 1998) (Fig. 2.1). The basin was connected to the extensive Neotethys Ocean that developed at the same time further to the north and west.

The Early Mesozoic rifting activity is considered by most authors to be the cause of the thinner crust identified within the Levant Basin (Fig. 2.1). The composition of the crust in the southeastern Mediterranean Sea is still under debate. Some researchers postulate an oceanic lithosphere that was formed when the initial rifting evolved into full scale sea-floor spreading (Makris et al., 1983; Ginzburg and Ben-Avraham, 1987; Ben-Avraham and Ginzburg, 1990). Others assume only stretched and intruded crust of continental origin that was formed during intra-plate rifting (Cohen et al., 1988, 1990; Hirsch et al., 1995; Gardosh and Druckman, 2006).

Rifting was followed by post-rift cooling and subsidence of the newly formed or modified lithosphere in the Levant Basin (Garfunkel and Derin, 1984; ten Brink 1987; Garfunkel, 1998). Starting from the Middle Jurassic a paleo-depositional hinge belt developed along the eastern margin of the basin. The hinge belt that was originally identified in deep wells along the present-day Israeli coastline (Gvirtzman and Klang, 1972; Bein and Gvirtzman, 1977) separated a vast shallow-marine platform on the east from a deep marine basin on the west. Several studies suggest that the Mesozoic shelf-edge extended near the entire southeastern Mediterranean coastline from northern Egypt to western Lebanon (Figs. 1.1, 2.1) (Walley, 1998, 2001).

The Late Cretaceous to Early Tertiary section is dominated by hemi-pelagic to pelagic sedimentation, marking the termination of the extensive, shallow-marine carbonate deposition on the Levant margin. The lithologic transition reflects a major drowning event and ecological changes associated with plate re-arrangement and the beginning of the closure of the Neotethys Ocean (Sass and Bein, 1982; Almogi-Labin et al., 1993).

The Late Cretaceous convergence of the Eurasian and Afro-Arabian plates initiated a northward-dipping subduction zone in the southern Neotethys Ocean, in the present-day areas of Cyprus and southern Turkey (Fig. 1.1) (Robertson, 1998). Far-field stresses related to the onset of subduction may have caused compression of the Levant passive margin and triggered the formation of anticlines and synclines throughout the Levant region, known as the ‘Syrian Arc’
(Krenkel, 1924) or 'Levantide' (Picard, 1959) fold belt. Well and seismic data suggest that the Syrian Arc deformation is generally associated with reactivation of Early Mesozoic normal faults in a reverse motion (Fig. 2.1) (Freund et al., 1975; Druckman et al., 1995). Evidence from Israel and Lebanon further indicates that the Syrian Arc deformation that started in the Late Cretaceous continued through the Tertiary (Eyal and Reches, 1983; Walley, 1998, 2001).

The continued plate convergence during the Oligocene and Early Miocene resulted in the formation of a subduction zone along the present Cypraen Arc plate boundary (Fig. 1.1)(Robertson, 1998). In the Levant region plate motion was associated with uplifting east of the Levant coast and updoming of the Arabian Shield south of it. This activity is reflected by a regional unconformity surface that is identified at the upper part of the Eocene strata, which separates the older carbonate section from the overlying Oligo-Miocene siliciclastics of the Saqyie Group (Picard, 1943; Ball and Ball, 1953). The deposition of Oligo-Miocene, coarse-grained clastic units is associated with widespread erosion and basinward transport of sediments through an extensive subaerial and submarine drainage system that developed on the Levant slope (Gvirtzman and Buchbinder, 1978; Druckman et al., 1995b).

A thick, Upper Miocene series of halite and anhydrite (Fig. 2.1) is associated with a huge sea-level drop and desiccation of the Mediterranean Sea known as the Messinian Salinity Crisis (Hsu et al., 1978, Gvirtzman and Buchbinder, 1978). The Messinian evaporites filled a 1-2 km deep topographic depression located west of the Levant coast (Tibor et al., 1992). Gradual sea-level rise during the Plio-Pleistocene resulted in renewed sediment transport and basinward progradation of a fine-grained siliciclastic wedge on the Levant shelf.

The breakup of the Arabian Craton by the end of the Oligocene brought into existence the Dead Sea plate boundary. This transform fault is several thousands of kilometers long, connecting the spreading center of the Red Sea with the Taurus collision zone in southeastern Turkey (Fig. 1.1)(Freund et al., 1970; Garfunkel, 1981). The Israeli segment is characterized by several, deep continental depressions, such as the Dead Sea basin, in which thick (up to 8-10 km) Miocene and Plio-Pleistocene sections accumulated (Fig. 2.1) (Kashai and Crocker, 1987; Gardosh et al., 1997).
3. Data Sets

3.1 Well Data

The well data used in this study are from 43 wells located within the basin or on its margin (Fig. 1.3). Stratigraphic subdivisions of these wells were taken from the lithostratigraphic data base of oil and gas wells drilled in Israel, compiled by Fleischer and Varshavsky (2002) (Table 3.1). Formation tops were used as control points in the construction of seismic depth maps (Table 3.1, Figs. 5.11-18). Synthetic seismograms and time-converted wireline logs of 14 wells, most of them located offshore and some near the coast, were used for correlation of well and seismic data (Table 3.1, Figs. 5.1, 5.2). The lithostratigraphy and biostratigraphy of selected wells were thoroughly reviewed and were used to construct two, detailed stratigraphic cross-sections (Figs. 4.5, 4.6). Additional well and outcrop data from the Levant margin were analyzed and were integrated into four composite sequence stratigraphic schemes of the Triassic, Jurassic, Cretaceous and Cenozoic (Figs. 4.1, 4.2, 4.3, 4.4).

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<th>Upper Miocene (Base Messin.)</th>
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<td>Green</td>
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<td>Cyan</td>
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<td>Violet</td>
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<td>Andromeda-1</td>
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<td>ASYM1</td>
<td>~7800</td>
<td>~2100</td>
<td>479/Yagur</td>
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<td>3900/Barnea</td>
<td>2234/TY</td>
<td>2040/Av</td>
<td>1303/BG</td>
<td>1120/Ziqim</td>
<td>1063/Mav</td>
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<td>3400/Barnea</td>
<td>2098/TY</td>
<td>1955/Av</td>
<td>1245/BG</td>
<td>1009/Ziqim</td>
<td>961/Mav</td>
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<td>Atlit-1 Deep</td>
<td>ATL11</td>
<td>~8500</td>
<td>1850/ in M. Haifa</td>
<td>45/Negba</td>
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<td>BYAM1</td>
<td>2432/TY</td>
<td>2197/Av</td>
<td>2087/BG</td>
<td>1479/AshCl</td>
<td>1479/NoVulc</td>
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<td>BRV1</td>
<td>2939/Daliyya</td>
<td>2197/Av</td>
<td>1627/BG</td>
<td>1380/Ziqim</td>
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<td>393/Ziqim</td>
<td>353/Mav</td>
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<td>132/BG</td>
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Time converted, used for calibration of seismic data
~xxxx= not penetrated, depth is estimated

Table 3.1- Summary of well data used in the present study. The depth of chronostratigraphic and lithostratigraphic tops in the wells is in meters below MSL. These depth values were used for calibration of seismic data and depth maps. Abbreviations of lithostratigraphic units are: TY- Talme Yafe Formation, Mt Sc-Mount Scopus Group, AV-Avedat Group, BG- Bet Guvrin Formation, AshCl- Ashdod Clastics, Mav- Mavqiim Formation.

3.2 Seismic Data

The seismic data set used in this research includes two, large marine surveys (A and B) that were carried out during the year 2001 in the framework of hydrocarbon exploration activity offshore Israel (Fig. 1.3). 98 multi-channel, 2D seismic reflection lines, totaling 12,000 km were interpreted (Fig. 1.3).
The A survey covers most of the Levant Basin from the Israeli shallow shelf to the submarine Eratosthenes Seamount, some 200 km northwest. The B survey extends in a more densely spaced grid from near the coastline to about 120 km offshore (Fig. 1.3). The long cable, small group interval and high-energy source used in the two surveys resulted in high-quality seismic data sets that display better vertical resolution and deeper penetration than older, offshore seismic vintages.

Two onshore Vibroseis lines, extending along the coastline from Haifa in the north to the Gaza area in the south, were also incorporated into the seismic data set (Fig. 1.3). Time-migrated profiles were interpreted on a seismic interpretation workstation using Paradigm-Epos3 software.

Four seismic lines of the A survey were reprocessed in the GII. The primary purpose of the reprocessing was to produce pre-stacked depth-migrated profiles and to extract seismic interval velocities that were used in depth conversion of time maps.

3.3 Gravity and Magnetic Data Sets

Gravity and magnetic data were acquired in the B seismic survey (Fig. 1.3). A total of 320,000 stations were measured along the seismic lines at intervals of about 20 m. The accuracy of the data (estimated on the crossing locations) is about a few nanoTesla for magnetic measurements and 0.1mGal for the gravity measurements.

The raw gravity data were filtered and Eotvos, Free Air and Bouguer corrections were applied. The IGRF was subtracted from the magnetic data. Both data sets were incorporated in the regional gravity and magnetic grids of the Levant area compiled by Rybakov et al, (1997) (Figs. 6.1, 6.2).

4. Stratigraphy

4.1 General

The Levant area is located on the northwestern margins of the Arabo-Nubian Shield. This shield was shaped during the Late Precambrian Pan-African Orogeny at about 850-700 Ma. It underwent cratonization at ~630-600 Ma and was uplifted and peneplained at 532 Ma (Garfunkel, 1980; Beyth and Heimann, 1999). The overlying sedimentary cover forms a northwest-thickening wedge, ranging from about 2 km in southern Israel through ca 7.5 km along the Mediterranean coastal plain and to ca 15 km in the eastern Mediterranean basin.

Throughout the Phanerozoic the Levant area formed a transitional zone between the continental domain in the southeast and the marine domain of the Tethys and Neotethys oceans.
to the west. The dominant sedimentary facies east and south of the present-day Mediterranean coast were thus epicontinental, consisting of alternating, shallow-marine carbonate, shale, and sandstone, with abundant coarse-grained components in the south. Slope and basinal sedimentation was dominant in the west, across the coastal plain and in the marine basin at times when this area subsided.

The Phanerozoic sedimentary section of the Levant is characterized by abundant coastal onlap-offlap cycles of different orders (Figs. 4.1 to 4.4). These cycles reflect recurring transgressions and regressions of the Tethys and Neotethys oceans onto the Arabian Craton. Relative high and low sea-level stands were controlled both by global eustatic changes and by local, vertical tectonic movements. The latter are evident by the substantial gaps in the sedimentary records in the Paleozoic, Early Jurassic and Early Cretaceous successions.

4.2 The Permo-Triassic Section

The earliest Phanerozoic sediments in the central and southern part of Israel are of Early Permian age (Weissbrod, 1969b, Eshet, 1990). Late Carboniferous to Early Permian uplift of the Negev and central Israel led to the erosion of the older Paleozoic section down to the Infracambrian Zenifim Formation (Garfunkel, 1988). Pre-Permian sections that escaped this erosion may have been locally preserved, e.g., the 300 m thick section of Cambrian sandstones and carbonates that are found in the southernmost tip of the country (Elat region; Weissbrod, 1969b). No Paleozoic section was penetrated in any wells in northern Israel.

The Permo-Triassic shallow-marine section accumulated in a large, N-S oriented embayment, roughly coinciding with today's Judean highlands. The section thickens gradually to the north and thins to the east and west (Weissbrod, 1969a; Druckman, 1974; Druckman and Kashai, 1981; Druckman et al., 1995a; Korngrin, 2004).

The western flank of this embayment is delineated by a 'basement high' that was penetrated in the Helez Deep-1, Gevim-1, and the Bessor-1 boreholes, located in the southern coastal plain (Fig. 4.5). Its position in the southern coastal plain may explain the thickening of the Permo-Triassic north and east of it. This 'basement high' was fractured and downfaulted to the west in several pulses, during the Late Triassic to Middle Jurassic times (Fig. 4.5). The Permo-Triassic stratigraphy was not penetrated, and is therefore unknown west of the 'basement high'. Gvirtzman and Weissbrod (1984) termed this feature the 'Helez High' and considered it as part of a regional 'geoanticline.'

The Lower Permian Sa'ad Formation overlies the Precambrian basement and Infracambrian section and is found in wells in the Negev area and southern coastal plain. Its northernmost occurrence is in the David-1 borehole, in central Israel. The Sa'ad Formation
consists of a 70-180 m thick fluvi-deltaic section (Weissbrod, 1969a). The deltaic sandstones may provide favorable reservoir properties.

The Upper Permian (Thuringian) consists of the Arqov Formation. The formation is found in wells in the Negev area and southern coastal plain where it comprises a 110-250 m thick section of shallow-marine carbonate, sandstone and shale. The unit has source, reservoir and seal potential. Noteworthy is its great petrographic similarities with the gas producing Khuff Formation in the Arabian Gulf area (Derin and Gerry, 1981; Weissbrod, 2005).

The Permian-Triassic boundary is marked by an hiatus of the Induan (lowermost Triassic stage). The boundary is marked by red clay, a characteristic fungal spike and reworked Lower Paleozoic microfossils (Eshet, 1990; Benjamini et al., 2005).

The Triassic section accumulated in a large, shallow-marine embayment oriented in an N-S direction. The section thickens from about 1000 m in the Negev to over 2600 m in the northern part of the country (the Devorah-2a well; Druckman and Kashai, 1981). The thickness of the Triassic section in the Helez Deep-1, Gevim-1, and Bessor-1 wells located on the ‘Helez High’ is reduced to a few tens to several hundred meters (Fig. 4.5) (Druckman, 1974, 1984; Druckman et al., 1994, 1995a); whereas the Gaash-2 and Atlit-1 deep wells located 70 and 130 km north of Helez Deep-1, respectively, penetrated over 1150 m of the Middle and Upper Triassic sections (Derin and Gerry, 1981, Korngrin, 2004). The thickness variations in the Triassic section reflect a regional northward dip of the basement and local vertical motions.

The Triassic section comprises a series of onlaps reflecting ten, stacked depositional cycles, which are described in the following sub-sections and are shown in Figure 4.1.

4.2.1 The early Olenekian cycle (Yamin Formation)

The Yamin Formation represents the first cycle. The cycle consists of some 120 to 155 m of nearshore and shallow-marine shale and carbonates and its time of deposition spans ~ 5 m.y.

4.2.2 The early Scythian to early Anisian cycle (Zafir Formation)

The Zafir Formation represents the second Triassic cycle. The cycle consists of some 175 to 320 m of shallow marine carbonate, calcareous shale and sandstone and its time of deposition spans ~ 3.5 m.y.

4.2.3 The early Anisian cycle (Ra'af Formation)

The Ra'af Formation represents the third cycle. This cycle consists of carbonate mudstones, packstones, and oolitic and oncolitic grainstones. Its thickness ranges from 80 m in the south to 170 m in northern Israel (Devora-2a well), and its time of deposition spans ~1.5 m.y.
Figure 4.1-Triassic stratigraphy and marine onlaps, Israel
Late diagenetic dolomitization affected the carbonates, mainly in the south. In the southernmost occurrence at Har Arif, the carbonates are sandy towards the top of the Ra'af Formation, probably representing the highstand progradation stage.

4.2.4 The late Anisian to early Ladinian cycle (Gevanim and Saharonim formations)

The lower member of the Gevanim Formation reflects a lowstand wedge. This clastic wedge, was deposited in the Negev area only within a time-span of less than 1 m.y. It is composed of fluvial to fluvio-deltaic coarse- to medium-grained sandstone and shale with abundant plant remains. The unit thins from 160 m in the Makhtesh Ramon area, to 25 m in the Sherif-1 well, 40 km northwards (Druckman, 1974). The thinning of the unit marks the onset of vertical movements related to the Early Mesozoic rifting (Freund et al., 1975).

Several tens of thousands of barrels of medium to high grade oil were produced from the lower Gevanim sandstones and the Raaf carbonates in the Zuk Tamrur-3 and Emmunah-1 wells, located in the western margins of the Dead Sea.

The Late Anisian - Early Ladinian middle and upper members of the Gevanim Formation mark the transgressive systems tract of the cycle. The maximum flooding stage was reached during the deposition of the burrowed carbonate mudstones topping the Fossiliferous Limestone Member at the lower part of the Saharonim Formation (found in outcrop in Makhtesh Ramon in southern Israel). The unit consists of fluvio-deltaic and nearshore sandstone and shale passing upward into alternating, meter thick, marlstone and storm beds of molluscan limestone. This part of the cycle spans ~3.5 m.y. (Druckman, 1974, 1976; Benjamini et al., 1993).

4.2.5 The Ladinian to early Carnian cycles (Saharonim Formation)

The middle and upper parts of the Saharonim Formation comprises the fifth, sixth and seventh depositional cycles. Their deposition lasted some ~3.8 m.y. These units consist of ammonite bearing limestones, shell beds, oncolitic and oolitic grainstones, stromatolite beds, burrowed mudstones, flat-pebble conglomerates, dolostones, marls and gypsum beds, all deposited in tidal to shallow marine environments (Druckman, 1969, 1976; Benjamini et al., 1993). The maximum flooding stage of each cycle is marked by a burrowed carbonate mudstone that can be traced throughout the entire Negev. The Saharonim Formation displays a significant thickness change, from 260 m in the south (Ramon-1 well) to 760 m in the northern part of Israel (Devorah-2 well).
4.2.6 The Carnian cycles (Mohilla Formation)

The Mohilla Formation comprises the eighth and ninth cycles. The lower boundary of the eighth cycle is identified by a sharp transition from the marine burrowed mudstone of the Saharonim Formation to the supratidal dolomicrites at the base of the lower member of the Mohilla Formation. The onset of the ninth cycle is marked by evaporite deposition in the upper member of the Mohilla Formation. The termination of the cycle is marked by the upper Carnian skeletal and stromatolite limestone.

The Mohilla Formation displays a substantial northward thickening from 200 m in the Ramon outcrop to 1000 m in the Devora-2a well (Druckman and Kashai, 1981). During its depositional time span (~ 5 m.y.) a marked differentiation in lithologies took place, i.e., thick accumulations of anhydrites (150 m in the Negev and 800 m in the northern part of Israel) were deposited in distinctive basins trending in a SW–NE direction. The basins were separated by dolomitic sections of reduced thickness (50 m in the Negev and 200 m in the north), indicating elevated swells between the basins. These facies and thickness changes indicate differential rates of subsidence apparently caused by vertical movements (Druckman, 1974; Druckman and Kashai, 1981). Normal faulting of several hundred meters during Carnian times resulted in the accumulation of the Erez Conglomerate on the foot-wall of the Helez fault (Druckman, 1984; Eshet, 1990). The Carnian vertical movements mark the second onset of extensional tectonics of the Early Mesozoic rifting phase.

4.2.10 The Norian cycle (Shefayim Formation)

The Shefayim Formation represents the last Triassic cycle. The unit was recognized only in the Atlit-1 and Gaash-2 wells in the northern coastal plain (Gvirtzman, 1981). The duration of the Shefayim cycle is estimated to be ~8.5 m.y. It is characterized by a reef facies consisting of sponges and other encrusting organisms, unidentified corals, and oolite beds (Derin et al., 1981; Korngrin and Benjamins, 2001; Korngrin, 2004). In the Atlit-1 well the Shefayim section reaches 1150 m (including about 270 m of Jurassic magmatic intrusions).

The Norian sequence was probably eroded in most of Israel during the latest Triassic to earliest Jurassic sea-level drop (Fig. 4.1). This global lowstand is marked by a notable unconformity surface of low relief, representing a matured peneplain surface. The hiatus between the Norian Shefayim Formation and the Pliensbachian Ardon Formation spans ~15 m.y. This unconformity is known throughout the Middle East. It is associated with complete exposure of Triassic terrain to karstic dissolution, laterization and the development of a thick paleosol referred to as the Mishhor Formation (Goldberg and Friedman, 1974).
4.3 The Jurassic Section

In southern Israel, the Jurassic succession consists of ~1200 m of continental to near-shore sandstone, shale and carbonate. It gradually thickens to over 3700 m of shelf carbonates in central Israel and thins to ~2200 m in northwestern Israel. South of Makhtesh Ramon the Jurassic section is entirely truncated by the Valanginian erosive unconformity. Significant thickness variations, mainly in the Lower Jurassic sections, have been related to normal faulting associated with rifting that eventually opened the Neotethys Ocean and established the continental margin of the Arabian Craton (see chapter 7). Accretion of reefs and carbonate platforms mark the shelf edge, located close to the present-day Mediterranean shoreline. In the offshore, more than 1,650 m of shelf, slope and basinal sediments accumulated.

The Jurassic section was partially penetrated in six wells only in the Mediterranean offshore: Delta-1A, Yam-2, Yam Yafo-1, Yam West-1, Foxtrot-1 and Asher Yam-1(Fig. 1.3, Table 3.1). Correlations between these wells and the onshore wells are displayed in the two cross sections (Figs. 4.5, 4.6) and in the seismic profiles. The biostratigraphy of the Jurassic section offshore has been a source of confusion: Different fossil groups used for dating the samples (e.g., pollen and spores, dinoflagelates versus benthic foraminifera) often produced contradicting results. Additional difficulties arose from the lack of core samples and from the occurrences of allochthonous rock constituents associated with mass-transport.

The description of the Jurassic succession presented in the following chapters is largely based on the well-sampled section found in outcrops and wells in the Negev area. The Jurassic section comprises a series of onlaps reflecting seven, stacked depositional cycles that are shown in Figure 4.2.

4.3.1 The Pliensbachian cycle (Ardon Formation)

The Ardon Formation is the earliest of the Jurassic cycles (Maync,1966; Derin and Reiss,1966; Goldberg and Friedman, 1974, Perelis-Grossowicz et al., 2000). The cycle overlies the Mishhor paleosol and consists of shallow marine, peritidal carbonates with minor amounts of evaporites. Its time of deposition spans ~ 5 m.y. (Buchbinder and leRoux, 1993). The unit thickens from a few tens of meters in Makhtesh Ramon to nearly 800 m in the Hilal well in northeastern Sinai, and almost 1000 m in the Helez Deep-1A well in the southern coastal plain.

These thickness variations are associated with an Early Jurassic faulting episode that occurred simultaneously with extensive volcanism found in the subsurface in the Carmel area. There, a thick 2500 m pile of basalts and pyroclastics, termed the Asher Volcanics, accumulated in a fault-controlled depression oriented in a NW-SE direction (Gvirtzman and Steinitz, 1983; Garfunkel, 1989; Gvirtzman et al. 1990). These volcanics have an intraplate basalt affinity.
Fig. 4.2 - Jurassic stratigraphy and marine onlaps, Israel. The blue bar denotes the Top Bathonian seismic horizon.
(Dvorkin and Kohn, 1989). The Asher Volcanics range in age from 206 to 189 Ma (Steinitz et al., 1983; Segev, 2000)

4.3.2 The Aalenian to early Bajocian cycle (Inmar Formation)

The Inmar Formation comprises the second Jurassic cycle (Perelis-Grossowicz et al., 2000). The Lower Member of the Inmar Formation consists of fluvial and shoreface sandstone with minor amounts of mudstone, passing upwards to the shallow-marine, inner-shelf carbonate of the Qeren Member. The latter marks the maximum flooding surface of the cycle. The Upper Member of the Inmar Formation consists of fluvial and fluvio-deltaic sandstone, siltstone and mudstone, marking a highstand, progradational systems tract. This part of the cycle passes into the Rosh-Pinna Formation (Derin, 1974), which represents the distal part of the highstand progradation wedge in the coastal plain and northern part of the country. In the coastal plain and central and northern Israel the Inmar and the Daya formations merge distally to form a ~1500 m thick carbonate platform of the Nirim Formation. This thick platform is built of mudstones, peletal and oolitic packstones, wackestones and grainstones. The unified cycle lasted ~5 m.y.

The Aalenian-Lower Bajocian sequence displays significant thickness variations in the Negev, ranging from 0 in the NE to over 900 m in the NW (Goldberg and Friedman, 1974; Druckman, 1977). These variations indicate the most significant vertical movement related to the Early Mesozoic rifting phase that is identified in southern.

4.3.3 The late Bajocian cycle (Daya and Barnea formations)

The Daya and Barnea formations form the third Jurassic cycle. The cycle begins with transgressive, shallow-marine carbonate and sandstone of the Daya Formation. The thickness of the unit changes gradually from ~100 m in Makhtesh Ramon to nearly 250 m in the coastal plain. In several wells near the present coastline the Daya Formation either passes laterally to or is overlain by the deeper-marine Barnea Formation. The latter unit consists of thinly bedded sponge rhaxes, spicules and radiolarian packstones, interpreted to reflect the maximum flooding conditions of the cycle. The duration of the Daya-Barnea cycle is ~5.5 m.y.

4.3.4 The Bathonian cycle (Sherif and Shederot formations)

The Sherif and Shederot formations and an 'Undivided' Middle Jurassic unit found in the offshore, comprises the fourth cycle. This cycle probably represents the smallest Jurassic onlap. In the Negev the unit consists of nearshore sandstones and peritidal and subtidal carbonates of the Sherif Formation. These may denote the lowstand to transgressive systems tracts of the cycle. In the central and northern coastal plain it comprises the oolitic shoals of the Shederot
Formation, which probably correspond to the highstand systems tract. The inland part of the Shederot-Sherif cycle thickens from ~100 m in the south to about 500 m in the coastal plain (Fig. 4.5).

In the offshore wells, the fourth Jurassic cycle is found only in the lower part of Yam West-1 (4829 m to T.D.) and Yam Yafo-1 (5033 m to T.D.). The presumed Bathonian succession in Yam West-1 consists of more than 421 m (bottom not reached at the T.D.) of oolitic pelletal and intraclast grainstones and skeletal grainstones. The grainstones are stacked in 20-50 m thick packages separated by a few meters of thick gray shale. On the Gamma Ray and Resistivity logs a total of 14 such packages were defined. At the base of the section three coarsening upward cycles (40-50 m thick) were detected. These are followed by five fining-upward (bell shaped), small-scale cycles (10-30 m thick). The upper part of the succession consists of five coarsening-upward cycles (25-40 m thick). The oolitic grainstones are interpreted as shoals, deposited in a platform setting that existed during the Bathonian in the Yam West-1 area.

In the Yam Yafo-1 well, the lowermost 752 m is dominated by siltstones and shale, interlayered with carbonates. The carbonate beds consist of skeletal and oolitic grainstones and packstones. Microconglomerates composed of all the above lithologies were detected in ditch samples. A 60 m thick mudstone interval rich in porifera spicules and rhaxes was recorded between 5260 and 5320 m. Following Conway (written comm., 2006) a Bathonian age is assigned to this interval. The lithologic characteristics are different from the Shederot Formation found in the Yam West-1 well or in the onshore. The strata in the lower part of the Yam Yafo-1 well are therefore considered as an 'Undivided' Middle Jurassic unit of a presumed Bathonian age. The existence of microconglomerates and the large amounts of clastics suggest that the 'Undivided' unit in Yam Yafo-1, unlike the Shederot Formation in Yam West, may represent an allochthonous section derived in part from the nearby shelf (Druckman et al., 1994; Gardosh, 2002).

The Bathonian unconformity marked by the Blue horizon is correlated to the top of the Shederot Formation in Yam West-1 and to the middle part of the 'Undivided' unit in the Yam Yafo-1 well. The age definition of these offshore sections and their relation to the Shederot cycle onshore is problematic due to contradicting biostratigraphic ages obtained from different fossil groups.

An age of Aalenian to Early Bajocian in the Yam West-1 well is indicated by the occurrence of the fossils *Gutnicella gr. cayexi* between 5080 and 5180 m and *Timidonella sarda* between 5180 and 5250 m. A similar age is indicated in the Yam Yafo-1 by the fossils “*Mesoendothyra*” sp. between 5338 and 5368 m and a badly preserved specimen of *Timidonella*
sarda at 5770 m (Perelis-Grossowicz, 2006, written comm.). In the Yam West-1 well Perelis-Grossowicz further inferred a Late Bajocian-Early Bathonian hiatus of ~6-7 m.y. at a depth of 4910 m. The ostracode *Fabanella ramonensis*, which is known from the Pliensbachian Ardon Formation in the Negev, was found at the very bottom of the Yam West-1 and Yam Yafo-1 wells together with *Gomphocythere* sp. known from the Bathonian of the Paris basin.

In contrast to the biostratigraphic ages described above, Conway (written comm., 2006) assigned an age of "not older than Bathonian" to the lower 200 m in the Yam West-1 and Yam Yafo-1 wells. Conway based his age determination on the occurrence of the dynocyst *Systematofora penicillata*.

The mixed occurrence (Pliensbachian and Aalenian –Bajocian) of foraminifera and ostracoda faunas in the Yam West-1 and Yam Yafo-1 wells suggestes that they are reworked in younger strata of Bathonian age, as indicated by the dynocyst assemblage. The correlation of the oolitic offshore interval in Yam West-1 with the onshore Bathonian Shederot Formation is further supported by the fact that the Bajocian Barnea Formation which underlies the Shederot Formation onshore was not encountered in the Yam West-1 well, as would be the case if the oolitic succession in this well were indeed of Aalenian –Early Bajocian age (Fig. 4.5).

4.3.5 The early Bathonian to Callovian cycle (Karmon, Zohar, Matmor and Brur formations)

The Karmon, Zohar, Matmor and Brur formations comprise the fifth cycle. The Karmon shale is a terrigeneous unit several tens of meters thick found in wells of the southern coastal plain below the Zohar limestone. It is interpreted as the lowstand systems tract of the cycle that was deposited following a sea-level drop and exposure of the underlying Shederot shelf (Fig. 4.5) (Gardosh, 2002). The base of the Karmon shale is an upper Bathonian unconformity surface that is correlated to the Blue seismic horizon (Fig. 4.2). The overlying Zohar and Matmor formations in the Negev area consist of shallow-marine, low-energy carbonate interlayered with minor amounts of shale and sandstone. In the coastal plain the equivalent strata are the oolitic, oncoid, skeletal grainstones of the Brur Formation. The three shallow-marine units are interpreted as the highstand systems tract of the cycle. The strata of the fifth cycle attain thicknesses of 160-180 m in the northern Negev and ~ 100 m in the coastal plain. An unconformity surface associated with karst phenomena found at the top of the Brur Formation marks the end of the cycle (Buchbinder, 1979).

In the offshore Yam-2, Yam West-1 and Yam Yafo-1 wells the Zohar Formation ranges in thickness from 119 to 138 m. It consists of oolitic, pelletal and intraclast and skeletal grainstones, mudstones and packstones made of poriferan spicules, and a few intercalations of
dark gray shale with radiolaria. The Zohar Formation conformably overlies the 'Undivided' Middle Jurassic unit in the Yam Yafo-1 borehole and the Bathonian Shederot Formation in the Yam West-1 borehole.

Based on a revision of the foraminiferal biostratigraphy in the three offshore wells, Perelis-Grossowicz (written comm., 2006) updated the age of the Zohar to uppermost Bathonian to lower Callovian. Conway (written comm., 2006) assigned a Callovian age, similar to the age assigned to the Zohar and Matmor formations onshore (Hirsch, 2005).

Over 2 Bcm of thermogenic gas was produced from fractured limestone of the Zohar Formation in the Dead Sea area (Zohar, Kidod and Qanaim fields). Sub-commercial quantities of high-grade oil were also recovered from oolitic grainstones of the Zohar Formation in the offshore Yam Yafo-1 and Yam-2 wells.

4.3.6 The Haifa Formation

The Haifa Formation is a thick and uniform carbonate succession of Bajocian to Callovian age (Derin, 1974) that is found in wells in the northern coastal plain and offshore area (Fig. 4.6). This succession is equivalent to the 3rd, 4th and 5th cycles of the central and southern parts of Israel. The Bathonian unconformity marked by the Blue horizon is correlated to the top of the Lower Haifa Member (Fig. 4.6)

4.3.7 The Oxfordian cycle (Kidod Formation)

The Kidod Formation forms the sixth depositional cycle. The unit consists of dark gray to black shale, with minor intercalations of carbonate mudstone and packstone. It ranges in thickness from 0 to 200 m and its deposition spans ~2.6 m.y. Its maximum thickness is found along a NE-SW trending belt, extending from Mt. Hermon towards the northwestern Negev. In most of the Negev the unit was entirely truncated by the Valanginian unconformity. In the northern coastal plain and northern Israel it is entirely absent. In the offshore Yam West-1, Yam-2 and Yam Yafo-1 wells the unit reappears and ranges in thickness from 170 to 340 m. In these wells it consists of gray, green and reddish shale and a few interlayers of poorly sorted grainstones containing intraclasts, pellets, forams, and echinoids. The Kidod shale is generally barren of fauna, though a few radiolaria, foraminifera, molluscs, echinoids and ostracodes were found. In the offshore the unit conformably overlies the Zohar Formation and underlies the Delta Formation. This predominantly transgressive unit marks the termination of the Middle Jurassic carbonate platform offshore Israel, while onshore, carbonate platform conditions were later reestablished.
An Oxfordian age was assigned to the Kidod Formation (Druckman et al., 1994; Gill et al., 1995; Conway, 2006, written comm.). Perelis-Grossowicz (2006, written comm.) attributed a Middle Callovian – Oxfordian age to the Kidod Formation, whereas Derin et al. (1990) assigned it an Oxfordian – Kimmeridgian age. In the Hermon and Maghara outcrops an early to middle Oxfordian age was assigned, based on ammonite zonation (Hirsch, 2005).

The distribution pattern of the Kidod Formation in the offshore wells and in central Israel and its absence along the coastal area may be attributed to submarine erosion or non deposition along the upper slope during late Oxfordian times. However, subaerial erosion cannot entirely be ruled out. This interpretation differs from that proposed by Derin, (1974), who related the absence of the Kidod Formation in the coastal plain to interfingering with carbonates of the Nir Am Reef. Derin (1974) further attributed the thick accumulation of the Kidod in the NE-SW trending belt to backreef lagoonal conditions. The proposed model herein indicates open marine conditions, which accords better with the rich ammonite fauna found in the Hermon and Maghara outcrops. The widespread distribution of Oxfordian shale in the Middle East better supports an open marine model for the Kidod shale rather than a shelf lagoon as proposed by Derin (1974).

4.3.8 The late Oxfordian to Tithonian cycle

The upper Oxfordian to Tithonian strata comprise the seventh cycle. This cycle, which lasted about 12 m.y., is assumed to represent the maximum onlap extension of all Jurassic cycles. It corresponds to the highest Jurassic global sea-level stand (Haq et al., 1988). The cycle comprises the Beer Sheva, Haluza, Nahal Saar, Haifa Bay, Bikfaya and Salima formations onshore Israel and southern Lebanon, and the Delta and Yam formations in the offshore wells. In the onshore it consists of 200 to 400 m of shallow marine carbonate mudstones, packstones, coral and stromatoporid reefs, and oolitic and oncolithic shoals (Nir Am, Be'er Sheva, Nahal Saar, Bikfaya formations). These carbonates are overlain by peritidal sandstones, siltstones and shales interlayered with low–energy, subtidal carbonates and numerous, locally developed oolitic shoals (in the Haluza and Salima formations). These Upper Jurassic units have escaped truncation by the Valanginian unconformity due to their low structural position (Eliezri, 1964).

In the offshore wells the Oxfordian to Tithonian sequence is represented by the Delta and Yam formations. The Delta Formation varies in thickness from 148 to 240 m. It consists of redeposited reefoidal limestones, oolitic and pelletal grainstones, skeletal grainstones, burrowed shale and minor amounts of fine to medium-grain, moderately sorted quartz sandstone. Twelve meters of the Delta Formation were cored in the Delta-1A well. The cores contain mainly burrowed black shale and rip-up clasts, with occasional 10-20 cm thick layers consisting of
centimeter-scale redeposited, oolitic and skeletal grainstones with fine to coarse lithoclasts. A few fining-upward cycles several centimeters thick were also observed. Three distinct volcanic bodies, 15-20m thick are found within the Delta Formation, in the Delta-1 well.

Friedman et al. (1971) interpreted a depositional slope environment for the Delta Formation. In the Yam-2 and Yam West-1 wells, three stacked fining-upward cycles are observed on the Gamma Ray and Resistivity logs that possibly represent submarine, migrating mid-fan channel fill. These fining-upward cycles are poorly developed in the Delta-1A well and absent in the Yam Yafo-1 well. An Oxfordian to Kimmeridgian age was assigned to the Delta Formation offshore (Derin et al., 1988, 1990; Moshkovitz, in Derin et al., 1988; Druckman et al., 1994; Gill et al., 1995; Perelis-Grossowicz and Conway, 2006, written comm.). The unit conformably overlies the Kidod Formation and underlies the Yam Formation.

The Yam Formation ranges in thickness from 100 to 387 m. It consists of dark, finely laminated shale and siltstone generally barren of faunal elements but occasionally containing some foraminifera and ostracodes, skeletal, oolithic and pelletal packstones and fine to coarse-grained grainstones, subangular to subrounded quartz sandstone, poorly sorted (Druckman et al., 1994; Gill et al., 1995). In the Delta-1 well volcanic fragments were also recorded (Derin et al., 1990). The Yam Formation conformably overlies the Delta Formation and underlies the Arsuf Formation. Its age was determined as Tithonian, however a Kimmeridgian age for its lower part is possible (Derin et al., 1988, 1990; Moshkovitz in Derin et al., 1988; Druckman et al., 1994; Gill et al., 1995; Perelis-Grossowicz and Conway, 2006, written comm.).

During the Late Jurassic a distinct shelf-edge was established separating the shallow shelf to the east from a slope and basin to the west. The Delta and Yam formations are thus interpreted as slope and basin deposits that were deposited during Late Jurassic lowstands. The absence of these units west of the shelf-edge is related to either: (a) non deposition on the Late Jurassic upper slope; or (b) erosion of both the upper slope and shelf deposits during the Valanginian unconformity.

Onshore, there is a hiatus of ~ 23 m.y. above the Upper Jurassic strata, from the Berriasian until the lowermost Aptian. Offshore, however, continuous sedimentation took place across the Jurassic-Cretaceous boundary. It is possible that Berriasian sediments were deposited onshore and were later eroded during latest Berriasian to Early Valangian sea-level drop (see next chapter).


<table>
<thead>
<tr>
<th>Formation</th>
<th>Delta 1A</th>
<th>Yam 2</th>
<th>Yam Yafo 1</th>
<th>Yam West 1</th>
</tr>
</thead>
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<tr>
<td></td>
<td>Top m.</td>
<td>Thickness m.</td>
<td>Top m.</td>
<td>Thickness m.</td>
</tr>
<tr>
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<td>200</td>
<td>4263</td>
<td>171</td>
</tr>
<tr>
<td>Yam</td>
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<td>100</td>
<td>4434</td>
<td>263</td>
</tr>
<tr>
<td>Delta</td>
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<td>218+</td>
<td>4697</td>
<td>198</td>
</tr>
<tr>
<td>Kidod</td>
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<td>340</td>
<td>4725</td>
<td>170</td>
</tr>
<tr>
<td>Zohar</td>
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<td>135+</td>
<td>4895</td>
<td>138</td>
</tr>
<tr>
<td>Shederot/Undefined</td>
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<td></td>
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<td>752+</td>
</tr>
<tr>
<td>Total Jurassic</td>
<td>518+</td>
<td>1107+</td>
<td>1658+</td>
<td></td>
</tr>
</tbody>
</table>

Table 4.1- Summary of Jurassic formation tops and thicknesses in offshore wells.

**4.4 The Cretaceous Section**

**4.4.1 Berriasian to early Valanginian cycle (Arsuf Formation)**

A global sea-level drop started in the Berriasian and culminated in the early Valanginian (Fig. 4.3) (Haq et al., 1988). At the same time, a large-scale uplift of the landmass took place, east and south of the present-day shoreline. This uplift was caused by heating related to an upwelling of a mantle plume associated with the Tayasir Volcanics and the shallow Ramon intrusions (Garfunkel and Derin, 1984; Gvirtzman et al., 1998). Consequently, intensive and widespread erosion took place. The intensity of the erosion increased from north to south, eventually resulting in the removal of most of the Phanerozoic section in the southern Negev (Eliezri, 1964; Weissbrod, 1969a; Druckman, 1974; Garfunkel and Derin, 1984).

The Arsuf Formation (a new formation name proposed herein) ranges in thickness from 63 to 200 m. (Table 4.1). It consists of carbonate debris interbedded with thin shale intercalations. The carbonate debris is apparently allochthonous, representing reworked carbonate platform sediments, e.g., skeletal, oolitic, pelletal, grainstone or packstone, occasionally sandy. The formation is characterized by a 'bow' shaped gamma ray pattern, reflecting a gradual decrease of mud content towards the crest of the 'bow' and an increase of mud again towards its top. This pattern characterizes waxing and waning of sedimentation on a mixed sand-mud fan (Emery and Myers, 1996). The sandy component of the Arsuf fan, however, is made of carbonate debris.

The age of the Arsuf Formation is latest Tithonian to Berriasian. However, an early Valanginian age cannot be ruled out (Fig. 4.3) (Derin et al., 1988, 1990; Moshkovitz, in Derin et al., 1988; Druckman et al., 1994; Gill et al., 1995). Reworked Jurassic faunal elements are
Fig. 4.3 - Cretaceous stratigraphy and marine onlaps, Israel.
Selected continental deposits are delineated by dotted lines
The green bar denotes Top Turonian seismic horizon.
occasionally present. This unit conformably overlies the Yam Formation and underlies the Gevar'am Formation. The redeposition of Jurassic microfauna in the Berriasian to early Valanginian sediments supports the interpretation that the main erosion phase took place during this lowstand (Druckman et al., 1994).

4.4.2 Valanginian to Aptian cycle (Gevar'am Formation)

The pronounced uplift and related sea-level drop of late Berriasian - early Valangian times resulted in intensive erosion and transportation of the erosional products towards the sea. The transportation conduit persisted along the sea floor, eventually resulting in the incision of a large submarine canyon (Gevar'am Canyon), which in the coastal plain area, reaches a width of ca 30 km and depth of ca 1000 m (Cohen, 1976). The canyon is filled by marine shales (Gevar'am Formation) of Valangian? to Barremian age with a few conglomerate and sandstone intercalations.

Sandstone intercalations up to 15-20 m thick were encountered in the Yam-2, Yam West-1 and Yam West-2 offshore wells. They are referred to as "Lower Sand" of Hauterivian age and "Upper Sand" of Barremian age (Fig. 4.3) (Isramco Report). Most of these intercalations display a serrated log motif, indicating sand/clay interchange, which characterizes a 'channel levee' system of proximal slope fan (Emery and Myers, 1996). The lower sand in the Yam West-1 well displays a blocky gamma-ray motif that probably represents a sand lobe that breached the channel's levee (crevasse splay). The interpretation the offshore sand beds further suggests that the basin floor should have been located westward of the Yam wells.

4.4.3 Hauterivian to early Aptian cycle (Helez and Telamim formations)

The (siliciclastic-rich) Helez Formation (up to 240 m thick) represents a gradual encroachment of sea onto the shelf, probably in Barremian times (Fig. 4.3). This rise culminated in the Lower Aptian and is reflected by the oolite-rich Telamim (270 m)/Ein el Assad (Muraille de Blanche) carbonate platform and the Zuweira marine intercalation within the largely continental Hatira sandstone of the Negev (Fig. 4.3) (Weissbrod et al., 1990; Lewy and Weissbrod, 1993). In the offshore, the Helez -Telamim cycle is represented by the Gevar'am Formation, especially by its upper part (Fig. 4.3). Inland, the base-level rise during deposition of the Helez Formation resulted in the proximal accretion of continental, fluvial deposits of the Kohal Formation (up to 400 m thick; Shilo and Wolf, 1989), the sandstones below the Nabi Said Formation in the Qiryat Shemone (Galilee) area and the Gres de Base or Chouf Formation (Walley, 1998) in Lebanon.
The time span of the Helez -Telamim deposition was ca 10 m.y., thus encompassing a second-order depositional cycle. The cycle terminates abruptly and is replaced by a pelagic shale interval (Yavneh Shale, ca 50 m thick) marked by the LC-3 electrolog marker (Fig. 4.3). This change reflects a drowning phase which hampered platform growth. In northern Israel the platform carbonate of the Ein el Assad Formation shows a pronounced deepening towards the uppermost part of the formation (Bachmann and Hirsch, 2006), presumably paralleling the L.C.-3 drowning phase. A global ecological event responsible for the demise and drowning of carbonate platforms in the Mediterranean area at the end of the lower Aptian (Castro and Ruiz-Ortiz, 1995; Wissler et al., 2003) could have been also responsible for the termination of the Telamim/Ein el Assad platform (Bachmann and Hirsch, 2006).

The overlying Hidra Formation in the Galilee proximal area begins with oolitic marls of open lagoon environment (Bachmann and Hirsch, 2006). It is portrayed as a late highstand progradation stage of the Telamim - Ein el Assad cycle (Fig. 4.3). In the middle part of the Hidra Formation, 10 m of recurring intercalations of fluvial sediments and ferruginous crusts signal a pronounced sequence boundary and lowstand systems tract deposits.

A notable sandstone horizon overlain by a dolomite bed was identified by Grader and Reiss (1958), directly above the Yavneh Shale. In their composite log of the Lower Cretaceous in the Helez area this horizon is termed the "Main Shale Break" (M.S.B. electrolog marker). A correlation is proposed between this sandstone bed and the coarse conglomerate at the base of the Aptian-Albian Talme Yafe-Yakhini cycles (see below). This lowstand event is also responsible for the prominent erosion surface, which the Talme Yafe sediments overlie.

4.4.4 Late Aptian cycle (upper Hidra to top Limestone Member of the Rama Formation)

The next cycle (in proximal areas of northern Israel and the Negev) includes the upper Hidra, the "Zumoffen" Limestone Member and the Deragot marine intercalation within the Hatira sandstone in the Negev (Fig. 4.3). Its time span is ca 7 m.y. Along the coastal Hinge Belt it is represented by a carbonate platform facies of the Yakhini Formation (Fig 4.3). West of the Hinge Belt it is represented by the Talme Yafe prism of carbonate debris. The Talme Yafe Formation unconformably overlies different units from the Aptian Telamim Formation to Middle Jurassic formations, with coarse conglomerate breccia at its base (Bein and Weiler, 1976). The reasons for this intense erosion are not clear. Unlike the truncation by the Gevar'am Formation, which was associated with a tectonic uplift, the truncation at the base of the Talme Yafe Formation could have been caused by the pronounced sea-level fall in middle Aptian times, which is clearly displayed in Haq's curve (Fig. 4.3).
4.4.5 Albian cycle

The restricted environment at the top of the "Zumoffen" Limestone Member of the lower part of the Rama Formation reflects a sequence boundary (Fig. 4.3) (Bachmann and Hirsch, 2006). The next cycle in proximal areas includes the Rama Formation (Couches a Knemiceras) of mixed platform and basinal carbonate and the Yagur Formation, representing the highstand-progradation stage (Fig. 4.3). A parallel section was also found in the Judea Mountains (Ramallah area; Shachnai, 1969). The time range of this cycle is ca 13 m.y. In the more proximal areas of the Negev, the well-known sandy glauconite zone just below the Late Albian carbonate of the Hevyon Member (Judea Group) signifies the maximum flooding interval of the Albian cycle. In the coastal plain and the offshore distal areas the sedimentation of the Yakhini and the Talme Yafe formations continued.

4.4.6 Cenomanian cycle

An erosional unconformity at the top of the Yagur Formation (Fig. 4.3) (Folkman, 1969; Kafri, 1986; Lipson-Benitah et al., 1997) is marked by conglomerates and breccia, calcrete and dedolomitization, indicating aerial exposure (Folkman, 1969) and is thus designated as a sequence boundary. Extensive volcanism products overlie this surface in the Carmel and Umm el Fahm (Segev et al., 2002). In the northern Negev, this surface truncates the rudistid buildup at the top of the Hevyon Member, which contains the foraminifer *Preaealveolina*, thus indicating an early Cenomanian age (Lewy and Weissbrod, 1993). The overlying cycle begins with a thin, ca 0.5 m glauconitic carbonate bank with *Pycnodonte vesiculosa* and with early Cenomanian ammonites, representing part of the transgressive systems tract of the En Yorqeam – Zafit subcycle (Fig. 4.3). In the Jerusalem and Hebron hills the top of the Kesalon Formation represents the same phase (Fig. 4.3) (Braun and Hirsch, 1994).

The substantial abnormal thickness (2700 m) of the Cenomanian Item Formation in the Item-1 well offshore (reexamination by S. Lipson Benitah, 2006) is attributed to a submarine canyon incised into the Albian sediments of the Talme Yafe Formation as a result of the Kesalon lowstand event (Fig. 4.3). Cenomanian, mixed planktic and benthic foraminifera were found in the Item Canyon fill. Notably, both the Kesalon event and the younger Mukhraqa channeling events (see below) are associated with volcanic eruptions in the Carmel, at the base of the Isfiye Chalk (=Negba Formation) and the Mukhraqa Formation (Segev et al., 2002).

This early Cenomanian erosional surface was also described in Jordan by Abed (1984) and by Schulze et al. (2003) as sequence boundary CeJo1 at the top of Member b of the Naur Formation. The time span of the emergence which caused the unconformity was brief and was
soon followed by extensive sea level rise (represented by the chalky Negba Formation), suppressing carbonate platform development, except in the Judean hills (partial drowning).

The strata of the Cenomanian cycle are mostly basinal, pelagic sediments such as chalccs with chert beds, e.g., the Deir Hanna Formation in the Galilee, the chalk complex of the Carmel and the Negba Formation of the coastal plain and the offshore areas (Fig. 4.3). This chalk event indicates a drowning phase suppressing and eventually replacing the Albian carbonate platform. Inland, the cycle terminates with the prograding platform carbonate of the Sakhnin, Weradim and Tamar formations. In the Negev, the cycle is divided into two transgressive-regressive cycles: (1) En Yorqeam/Zafit, (2) Avnon/Tamar (Fig. 4.3).

Bet Oren– Atlit incised canyon of Middle/Late Cenomanian age- In the Carmel region, pelagic distal ramp deposits of the upper Cenomanian (Junediya Formation) are overlain by the lower part of the Late Cenomanian Mukhraqa Formation (age emendation by Buchbinder et al., 2000). Various types of transport phenomena characterize the latter unit, and are interpreted as distal bioclastic material shed off the prograding highstand deposits of the Late Cenomanian. ‘Reefal’ sediments of this unit were reported in an E-W trending erosive channel 10 km long which breaches the shelf edge near Atlit (Bein, 1977, figs. 2, 3). This channel may have been initiated as a large-scale slope failure. The erosion cuts down to early Cenomanian sediments. The age of the "reefal" Mukhraqa facies was emended to Late Cenomanian (instead of Turonian), based on the presence of caprinids (Buchbinder et al., 2000). The time of incision of this channel should, therefore, be late Middle Cenomanian to pre or early Late Cenomanian. Presumably, it is related to Haq's (1988) major sequence boundary Ce3/Ce4 at 94 m.y. (recalibrated to Gradstein et al., 1995) following a 60 m sea-level fall.

4.4.7 First Turonian cycle

This cycle begins with a major Late Cenomanian drowning phase which resulted in the demise of upper Cenomanian platforms (Tamar, Weradim, Sakhnin) followed by different time durations of condensation and hiatuses (Fig. 4.3) (Buchbinder et al., 2000). Sedimentation was later resumed in hemi-pelagic facies (Fig. 4.3) (Ora, Yirqa and Daliyya) and finally carbonate platforms were reestablished in proximal areas: Shivta; "Vroman" (Fig. 4.3). The cycle was abruptly terminated by a pronounced sea-level fall, resulting in the exposure of the shelf and the development of a system of rivers and overbanks, depositing quartzose sandstone and shale of the "Clastic Unit" (Fig. 4.3) (Sandler, 1996). In distal areas (coastal plain and Mediterranean offshore) basinal sediments of the Daliyya Formation either accumulated, or were replaced by condensed sections and hiatuses.
4.4.8 Middle Turonian to late Coniacian cycle (upper Bina, Gerofit and Zihor formations)

The cycle begins in proximal areas with platform carbonate of the Gerofit and upper Nezer formations (Fig. 4.3). A deepening trend developed towards Early Coniacian times and in late Coniacian the platform facies was abruptly replaced by pelagic-rich beds.

4.4.9 The Santonian to Paleocene basinal sediments (Mount Scopus Group)

Basinal-type sedimentation started in the late Coniacian and persisted inland throughout the Santonian to Paleocene times. These sediments are represented by the Mount Scopus Group (Flexer, 1968) and are dominated by pelagic chalk both in far proximal areas and in the offshore. They contain also marl, chert and phosphate. Sedimentation took place in the framework of an upwelling system that affected the southern coasts of the Tethys Ocean along a belt encompassing today's North Africa and the east Levant margins (Almogi-Labin et al., 1993). The widespread pelagic sedimentation reflects a major drowning of the shallow-marine Judean platform that occurred due to oceanographic, ecological change, probably as a consequence of re-arrangement of the plates bordering the Mediterranean Tethys. Several eustatic lowstands at that time (Flexer and Honigstein, 1984; Gardosh, 2002) did not effect the extensive penetration of the sea far inland and thus could not be expressed in the onlap curve (Fig. 4.3).

The Mount Scopus section contains organic-rich carbonate intervals. These bituminous rocks were deposited in the highly productive environment of small morphotectonic basins (Syrian Arc type synclines) and contain up to 20% of Type II kerogen (Tannenbaum and Aizenshtat, 1985; Gardosh et al., 1997). The Senonian bituminous rocks show geochemical correlation to asphalt, oil and gas accumulations found near the Dead Sea Lake and are considered to be the source of hydrocarbons that were generated within the Dead Sea Basin (Tannenbaum and Aizenshtat, 1985; Gardosh et al., 1997).

4.5 The Tertiary Section

4.5.1 General

Facies patterns of the Lower and Middle Eocene are predominantly pelagic. Although seemingly they contain shallower neritic interludes of nummolidic carbonates, the mixture with pelagic elements indicates a mass transport origin in a deeper basinal setting (Benjamini, 1993) beginning in latest Early Eocene times (Buchbinder et al., 1988).

Remarkably, the drowning conditions that started in the Sennonian persisted during most of the Tertiary; for about 70 m.y. and a true carbonate platform was re-established for a brief time span (ca 1 m.y.) only in Middle Miocene times (the Langhian Ziqlag Formation).
<table>
<thead>
<tr>
<th>SERIES</th>
<th>STAGES</th>
<th>BIOSTRATIGRAPHY</th>
<th>Shoreline</th>
<th>Judean Hills</th>
<th>SEA LEVEL</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pliocene</td>
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<td>7.12</td>
<td>Gr. humerosa</td>
<td>Slumped complex</td>
<td>Miller et al. 2005</td>
</tr>
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<td>N16</td>
<td>Gr. acostaensis</td>
<td>Turbidite sands</td>
<td></td>
</tr>
<tr>
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<td>N10</td>
<td>Gs. ruber</td>
<td>Mid. Ziqim</td>
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<tr>
<td>Aquitanian</td>
<td>N4</td>
<td>Pr. kugleri</td>
<td>Lower Ziqim</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Chattian</td>
<td>N2</td>
<td>Gg. ciperoensis ciperoensis</td>
<td>Cgl + sandstones, offshore</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rupelian</td>
<td>P19</td>
<td>Gg. ampliapertura</td>
<td>Mavqim Evap.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Priabonian</td>
<td>P16</td>
<td>Gr. ceroazulensis</td>
<td>Pattish Reef</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bartonian</td>
<td>P15</td>
<td>Gk. semiinvoluta</td>
<td>Canyons’ incision</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lutetian</td>
<td>P13</td>
<td>O. beckmanni</td>
<td>Bet Guvrin</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Fig. 4.4 - Tertiary stratigraphy and marine onlaps, Israel. The colored bars denote seismic horizon.
4.5.2 Late Middle Eocene to Late Eocene cycle

In proximal areas of the northern Negev a major unconformity was found between the Middle Eocene chalks of the Maresha Formation and the hemi-pelagic marls of the Qeziot Formation (Fig. 4.4). Benjamini (1979, 1984) indicated that the maximum hiatus between the two formations spans the P10 to P14 biostratigraphic zones. The Qeziot and the Bet Guvrin (its equivalent in the coastal plain and the Shefela areas, Fig. 4.4) formations are included in the Saqiye Group (Gvirtzman, 1970). The latter signify the "Upper Clastic division of Ball and Ball (1953). Occasional mass transported conglomerates are found especially in distal areas. In the Shefela, the Bet Guvrin Formation is of the Upper Eocene semiinvoluta zone (P15). The latest Eocene (P16) zone is missing. Siman-Tov (1991) reported on the rare occurrence of the latest Eocene T. cerroazulensis zone in Ofaqim and in Tel-Nagila. In other Upper Eocene locations this zone was truncated by the following major Early Oligocene unconformity. The Har Aqrav limestone (Fig. 4.4) which overlies the Qeziot Formation in the western Negev (Benjamini, 1980) apparently represents the highstand progradation phase of the Late Eocene cycle. Markedly, the hemi-pelagic facies of the Late Eocene is primarily uniform, both inland and far offshore.

Unlike other locations, the Eocene succession in the synclinal area of Ramat Hagolan (Michelson and Lipson-Benitah, 1986) shows an uninterrupted Early to Late Eocene biostratigraphic succession from P9 to P15. Therefore, Michelson and Lipson-Benitah (1986) regarded the Upper Eocene (P15) chalk of the Ramat Hagolan as the Maresha Formation (Fig. 4.4).

4.5.3 Latest Eocene - earliest Oligocene erosion phase

A major unconformity is found between the Late Eocene and Early Oligocene depositional cycles (Fig. 4.4). It could have been related to eustatic sea-level fall at 34 m.y. (shown by Haq et al., 1988 and Miller et al., 2005). Local uplift movements associated with Syrian Arc contraction (see chapter 7) or with doming of the Arabian Shield prior to the rifting of the Red Sea (Garfunkel, 1988) further intensified the erosional processes. A regional unconformity surface (peneplain) developed in the Negev and the Judea Mountains (Picard, 1951; Garfunkel and Horowitz, 1966). Zilberman (1992, 1993) named this feature “the upper erosion surface” and pointed out that it postdates Upper Eocene deposits and predates Miocene continental deposits of the Hazeva Formation. The presence of this erosion surface in the Judean hills indicates that the anticlinorial backbone of the country had already emerged at that time.
Buchbinder et al. (2005) matched the first incision of canyons on the Levant margin to the Early/Late Oligocene boundary (Fig. 4.4). Druckman et al. (1995b) matched it to the pre P19 – *ampliapertura* Zone (Fig. 4.4). The main incision in the Ashdod-1 well (Buchbinder et al., 2005, fig. 7) may be located at the bottom of the "Lower Sands", i.e., either between P19/P20, or within the P20, P21 zones (Fig. 4.4). Alternatively, it may be put at the base of the Lower Oligocene (P18) conglomerates. Obviously, the biostratigraphic resolution is not refined enough to determine the precise age of this erosion event because of intense contamination and redeposition in the cutting samples. However, for the sake of convenience and simplicity in interpretation of the seismic data the initial phase of the canyon's incision was placed at the base of the Oligocene sedimentary package, and this surface is represented as the Red horizon in the seismic profiles (Fig. 4.4, see chapter 5).

4.5.4 Early Oligocene cycle

Patchy outcrops of lowermost Oligocene sediments (Zone P18, Fig. 4.4) are scarce and are sparsely spread from the Golan to the northern Negev. They are more common in boreholes along the Mediterranean coastal plain and offshore areas. They invariably consist of deepwater pelagic, marly chalks, and their aerial distribution largely follows the distribution of the Upper Eocene deposits, except for their absence in the southern Negev (Buchbinder et al., 2005). They thus indicate that most of Israel except the southern and central Negev was covered by sea in earliest Oligocene times.

4.5.5. Late Rupelian-earlyChattian lowstand wedge and transported clastics

Sediments of P19-P21 zones (*ampliapertura* to *opima opima*, Fig. 4.4) have a limited inland distribution, especially the P19 Zone. Both are of hemi-pelagic facies, however, sediments of the *opima opima* Zone penetrate more inland to the Lower Shefela region (Lakhish Formation), displaying slumped blocks, debris flows and occasional sandy calcareous turbidites (Lakhish outcrops). In the Lakhish Formation the mass transported sediments are covered by a few meters of in situ large-foraminifera limestone representing a lowstand prograding wedge (Buchbinder et al., 2005). More than 300 m of sand and conglomerates of this age were penetrated in the Ashdod wells in the coastal plain. It is most likely that coarse-grained clastics could have been transported several tens of kilometers further to the west into the Levant Basin.
4.5.6 Late Oligocene (P22) – Early Miocene (N4) cycles

These cycles represent the last major inland transgressions (Fig. 4.4). They are separated by an erosional unconformity of moderate magnitude. The onlap of the Early Miocene cycle reached across the rift valley to the Golan area (Fig. 4.4, upper Ha'on Member, Buchbinder et al., 2005). However, their hemi-pelagic sediments were largely eroded in the hilly backbone of the country in later periods. The facies is ubiquitously pelagic or hemi-pelagic (Bet Guvrin type), except in the most proximal outcrop in the Golan.

4.5.7 The Burdigalian lowstand (N5-N7)

Sediments of this interval are limited to the coastal plain and offshore area. Canyon incision was renewed and mass transported deposits occasionally accumulated, e.g., the Gaza sands (Fig. 4.4; Martinotti, 1973).

4.5.8 Early Middle Miocene cycle (Ziqlag Formation)

This cycle is unique in the sense that it shows a development of a carbonate platform (Ziqlag Formation, Fig. 4.4) unconformably abutting the western flank of the Judean anticline, in times of dominant muddy deposition (Buchbinder et al., 1993; Buchbinder and Zilberman, 1997). However, its life span was short (ca 1.5 m.y.). The high sea level at that time facilitated the development of the carbonate platform by curbing siliciclastic supply. Hemi-pelagic deposition of the Bet Guvrin facies, however, persisted in the basin (Fig. 4.4). On the other hand, a thick (250 m) succession of conglomerate, consisting of limestone, chalk and chert pebbles apparently derived from Cretaceous to Tertiary formations, was found in the Nahal Oz-1 well in the Afiq Canyon (Druckman et al., 1995b). Its distribution probably spread towards the offshore continuation of the Gaza Canyon.

4.5.9 The Middle Miocene Seravallian crisis

The Ziglag episode terminated abruptly when the sea level dropped and erosion and canyon incision resumed, leaving behind their products: conglomerates and overbank deposits (Bet Nir conglomerate, Buchbinder et al., 1986). This happened despite the relatively high global sea level. Apparently there is evidence for a climatic, ecological and salinity crisis in the Mediterranean during Serravallian times, and as a result, the Mediterranean sea-level curve departed from that of the global sea level (see Martinotti, 1990; Buchbinder et al., 1993). This lowstand episode produced a distinct seismic reflector in the basin, namely the Cyan horizon (Fig. 4.4, chapter 5). Small onlap episodes of limited landward penetration are indicated by the lower and middle Ziqim "mini" cycles (Fig. 4.4).
4.5.10 Late Miocene – Tortonian to Messinian cycle (N16-N17)

Although of limited inland penetration, sediments of this cycle are relatively common in outcrops (Buchbinder and Zilberman, 1997) and especially in boreholes in the coastal plain (e.g., the Ashdod wells, Fig. 4.4). The cycle is marked by the development of coral (mostly *Porites*) and algal reefs of the Pattish Formation (Fig. 4.4). However, in the offshore extension of the Afiq (Gaza) Canyon, hemi-pelagic muds are found, interbedded with mass transported sands, conglomerates of chert pebbles.

4.5.11 The Messinian Mavqiim cycle

The Messinian cycle is composed of a thick evaporitic series found throughout the Mediterranean region (Hsu et al., 1973a,b; Neev et al., 1976; Ryan, 1978). The deposition of Mavqiim evaporites (Fig. 4.4), which are composed mainly of halite and some anhydrite, followed a major sea-level drop associated with the closure and desiccation of the Mediterranean Sea (Hsu et al.1973a,b; Hsu et al., 1978, Gvirtzman and Buchbinder, 1978). The area of evaporite accumulation extended throughout the Levant basin to about 20-40 km west of the present-day coastline (Gardosh and Druckman, 2006). The Messinian evaporites most likely filled a pre-existing topographic depression (Tibor et al., 1992) and are not associated with significant, local tectonic movements. However, during short highstand episodes the brine covered the slope and penetrated through deep canyons further inland. Remnants of the Messinian brine in the form of a few tens of meters of thick halite and anhydrite beds were encountered in various wells in the coastal areas of Israel (Druckman et al., 1995b). On the seismic data the base of the Messinian cycle is marked by the Purple horizon (Fig. 4.4).

4.5.12 Plio-Pleistocene cycles

The abrupt sea-level rise in the Early Pliocene terminated the evaporite deposition and resulted in the deposition of hemi-pelagic clays and marls of the Plio-Pleistocene Yafo Formation (Fig. 4.4) reaching over 1000 m in thickness. The unconformity surface at the top of the Mavqiim evaporites and the base of the Yafo Formation is marked on the seismic data by the Violet marker (also known as the M seismic marker; Neev et al. 1976) (Fig. 4.4, chapter 5). Sedimentation during the lowermost Pliocene succession was highly condensed. After deposition of a few tens of meters of hemi-pelagic marl in a period of ca 1 m.y., a significant sea-level drop occurred, estimated to correspond with the 4.37 m.y. sequence boundary of Hardenbol et al. (1998) (correlated with Haq et al.'s (1988) 4.2 m.y. sequence boundary). This lowstand resulted in the deposition of gas-bearing sands of turbidite origin in a basin floor or lower slope setting (Druckman, 2001; Oats, 2001). On land, this lowstand event is correlated...
with a boulder conglomerate separating between the Pliocene sequences I and II of Buchbinder and Zilberman (1997) in the Beer Sheva area.

Hemi-pelagic deposition was resumed during the following highstand and the rate of sedimentation gradually increased, eventually resulting in a progradational topset-clinoform pattern, clearly portrayed in the seismic sections (Figs. 5.1, 5.3).

The next significant events expressed in the seismic records are chaotic intervals of slump deposits of Late Pliocene and Pleistocene times (Frey-Martinez et al., 2005). Most notable is the large scale-slumping zone, ca 150 m thick, of Late Pliocene times, which may also be related to a lowstand phase.

5. Interpretation of Seismic Data

5.1 Seismic Horizons

The regional, 2D seismic profiles of the A and B surveys show a continuous series of seismic reflection that is 5 to 6 sec thick near the Israeli coast and reaches up to 10 sec (15-17 km) in the far offshore (Figs. 5.1, 5.3, 5.8, 5.9). This series comprises the entire sedimentary rock section that accumulated and was preserved within the Levant Basin during the Phanerozoic. The Phanerozoic basin-fill is subdivided into eight distinct seismic units that are separated by seven seismic horizons (Fig. 5.1, Table 5.1). Each of these horizons is interpreted as an unconformity surface that bounds a regional, basin-scale depositional unit. All the interpreted surfaces mark significant changes in seismic character within the basin-fill (Fig. 5.1). Some are recognized also by truncation or onlapping and downlapping reflections (Figs. 5.1, 5.3). The interpreted seismic markers extend throughout the entire basin; therefore they are not correlated to individual seismic reflections but rather to the tops (or bottoms) of separable seismic packages.

Well control on the seismic interpretation is possible only in the relatively shallow, eastern part of the basin. In this area the horizons show good fit to the tops of lithostratigraphic units identified in the wells (Figs. 5.1, 5.2). The correlation of seismic units from the shallow to the deep part of the basin, particularly for the deeper Brown, Blue, Green and Red horizons, is hampered by extensive faulting and tilting and is therefore somewhat speculative.
The lithologic column below the TD of the Yam West-1 well is inferred.

The Yam-2 and Yam-West-1 wells in the southeastern part of the basin.

**Fig. 5.1** Interpreted seismic profile showing the correlation of seismic horizons to

- Senonian-Eocene
- Plio-Pleistocene
- Messinian
- M.-U. Miocene
- M.-U. Jurassic-Turonian
- Precambrian (?)
- Crystalline Basement

---

**Legend:**
- **TWT (ms)**
- **Yam West-1 (appr. 2 km SW)**
- **Yam-2**
- **E**
- **W**
Fig 5.2 - Synthetic seismograms of Yam West-1 and Yam Yafo-1 showing the correlation of the seismic reflections to lithostratigraphic tops in the wells. The synthetic seismograms were prepared from acoustic (AC.) and density logs, and a zero-phase, normal 25 Hz Ricker wavelet.
Table 5.1- Correlation of interpreted seismic horizons to chrono- and lithostratigraphic unit tops.

<table>
<thead>
<tr>
<th>Seismic Horizon</th>
<th>Chronostratigraphic Top</th>
<th>Lithostratigraphic Top</th>
</tr>
</thead>
<tbody>
<tr>
<td>Violet</td>
<td>Late Messinian (Late Miocene)</td>
<td>Mavqiim (M horizon)</td>
</tr>
<tr>
<td>Purple</td>
<td>Early Messinian (Late Miocene)</td>
<td>Ziqim/Ziqlag/AshdodClst.(N horizon)</td>
</tr>
<tr>
<td>Cyan</td>
<td>Langhian (early Middle Miocene)</td>
<td>Bet Guvrin/Lakhish</td>
</tr>
<tr>
<td>Red</td>
<td>Eocene</td>
<td>Avedat</td>
</tr>
<tr>
<td>Green</td>
<td>Albian to Turonian</td>
<td>Talme Yafe/Yagur/Negba/Daliyya</td>
</tr>
<tr>
<td>Blue</td>
<td>Bathonian</td>
<td>Shederot/ within 'undivided' M. Jurassic</td>
</tr>
<tr>
<td>Brown</td>
<td>Precambrian ?</td>
<td>Crystalline basement</td>
</tr>
</tbody>
</table>

5.1.1 The Brown horizon

The brown horizon is the deepest seismic marker interpreted in this study. It is correlated throughout the basin with the transition from a chaotic and reflection-free seismic character below to a continuous, parallel to divergent, high-amplitude reflection series above (Fig. 5.1). No direct well control is possible for this deep marker in the offshore. We interpret the transition as characterizing a major acoustic boundary between non-layered, magmatic and metamorphic complexes and the overlying, layered Paleozoic-Mesozoic sedimentary interval. The Brown horizon is therefore considered as the 'near top of the crystalline basement' of an assumed Precambrian age.

The transition in seismic character is more evident in the southeastern corner of the study area, near the Yam West-1 well, where it is found at a relatively shallow depth of about 5-6 sec (Figs. 5.1, 5.3). In this area the seismic boundary possibly correlates to the top of the Precambrian to Infrcambrian basement rocks, penetrated by the Helez Deep-1 and the Gevim-1 wells, located about 15 km east of the coastline (Fig. 1.3). The resolution of the seismic data in the offshore generally decreases below 7-8 sec due the reduction of the seismic energy. However, the transition between the chaotic seismic character and the overlying, high-amplitude partly continuous reflection series is identifiable at the bottom of many seismic profiles that cross the central, deep part of the basin (Figs. 5.8, 5.9).

5.1.2 The Blue horizon

The Blue horizon is correlated to a distinct change in seismic character from high-amplitude; continuous reflection series below to discontinuous, low-amplitude occasionally shingled reflections above (Fig. 5.1 at 4500 ms between Yam-2 and Yam West-1 wells). The change in seismic character is most evident in the southeastern area, though it is observed also in
Fig 5.3- Interpreted, E-W oriented seismic profile through the Yam-2 and Yam West-1 wells, showing the area of transition from the elevated, Mesozoic thrust belt on the east to the deep, Tertiary basin on the west.
Fig 5.4- Interpreted composite seismic profile, generally oriented in a N-S direction along the shallow part of the basin, showing the continuity of seismic horizons from Yam West-1 (south) to Yam Yafo-1 (north).
Fig 5.5- Interpreted seismic profile, oriented in a NE-SW direction showing the correlation of seismic horizons in the deep part of the basin.
the central and western parts of the basin where the horizon is found at the top of a relatively continuous high-amplitude seismic event below less continuous and lower amplitude reflections (Figs. 5.3, 5.4, 5.5, 5.8, 5.9, 5.11).

The correlation of the Blue horizon from Yam West-1 to Yam Yafo-1 presents some difficulty. At Yam West-1 well it is correlated to a high-amplitude event found at the lower part of the well near the top of the Bathonian Shederot Formation (Fig. 5.2). At the Yam Yafo-1 well it is correlated to a high-amplitude event within an undivided Jurassic section of poorly defined age (probably Aalenian to Bathonian) (Fig. 5.2). Furthermore, there is a significant difference between the two wells in the lithology of the Middle Jurassic section. It is therefore unclear whether the seismic event represents the same chronostratigraphic unit in this area.

The change in character below and above the Blue horizon is, however, identifiable and most likely signifies a basin-wide seismic boundary. According to our stratigraphic analysis of the Jurassic section (chapter 4.3) the Blue horizon may correlate to the transition from a Lower Jurassic, shallow-marine carbonate section to a Middle-Upper Jurassic deep-marine section that was deposited during the passive margin stage.

5.1.3 The Green horizon

The Green horizon corresponds to the top of a distinct high-amplitude reflection package below lower-amplitude reflections, chaotic and reflection-free zones (Figs. 5.1 to 5.8). At the eastern part of the basin the marker is an onlapping surface (Fig. 5.1). Well control shows that the Green horizon correlates to the tops of several Middle Cretaceous units. These are the Albian Talme Yafe and Yagur formations, the Cenomanian Negba Formation and the Turonian Daliyya Formation (Fig. 5.2) (See also Table 3.1). In most wells of the study area the marker is overlain by strata of the Upper Cretaceous Mount Scopus Group (Figs. 5.1, 5.2, Table 3.1).

Throughout inland Israel a Middle Cretaceous unconformity is evident by a depositional hiatus and an abrupt lithologic change between shelf and slope carbonates of Albian to Turonian age (Judea and Talme Yafe formations) and pelagic marl and chalk of Senonian to Paleogene age (Mount Scopus and Hashefela groups) (Gvirtzman and Reiss 1965; Flexer 1968, Flexer et al., 1986). The change in seismic character corresponds to this well-recognized regional surface that is associated with drowning of the Middle Cretaceous carbonate platform (Sass and Bein, 1982; Almogi-Labin et al., 1993).

The correlation of the shallow, eastern part of the basin to the deep, western part requires a certain degree of jump-correlation across conspicuous sets of reverse faults (Figs. 5.3, 5.8). No well control is available in the far offshore. However, the identification of the Green marker at the top of a continuous, high-amplitude reflection series in the deep basin (Fig. 5.5) is well
Fig 5.6- Interpreted seismic profile showing the transition area from the elevated Mesozoic thrust belt to the deep Tertiary basin. The Red horizon is a base Oligocene unconformity (correlated to the Yam West-2 well) that truncates the Senonian-Eocene section at the tops of Syrian Arc fold structures.
In the Hannah-1 well the Oligo-Miocene fill overlies the Lower Cretaceous Gevar'am Formation. A base Oligocene unconformity truncates the Cretaceous, Senonian, and Eocene section. The Red horizon is a base Oligocene unconformity that truncates the Cretaceous, Senonian, and Eocene section. Fig 5.7 - Interpreted composite seismic profile through the Hannah-1 well. The Red horizon is a base Oligocene unconformity that truncates the Cretaceous, Senonian, and Eocene section.
Fig 5.8- Interpreted, NE-SW regional seismic profile showing extensional and contractional structures in the central part of the basin.
supported by the overall continuous seismic character of this unit as well as by the assumed vertical offsets on the reverse faults.

5.1.4 The Red horizon

The Red horizon is an unconformity surface that is characterized throughout the eastern part of the basin by truncation, incision and onlapping (Figs. 5.3, 5.6, 5.7). Well control shows that the Red horizon correlates to an Early Oligocene unconformity at the top of the Eocene Avedat Group or older stratigraphic units (Fig. 5.2, Table 3.1). In most of the wells of the study area the marker is overlain by strata of the Oligo-Miocene Saqiye Group.

The correlation of seismic profiles to the Yam West-2 and Hannah-1 well further demonstrate the amount of erosion associated with this seismic marker. In Yam West-1 the Red horizon is found at the top of an eroded, 30 m thick Eocene section (Fig. 5.6). At the Hannah-1 well the marker is correlated to the base of a deep canyon, incised within the Lower Cretaceous Gevar'am Formation, filled with Oligocene to Lower Miocene strata (Fig. 5.7). From these two wells the marker is carried across the top of several eroded structures down to the deep basin, where it merges with a series of high-amplitude, continuous seismic reflections at a depth of about 5-6 sec (Figs. 5.6, 5.7).

It should be noted that the present interpretation of the Red horizon within the deep basin is a revision of the results presented by Gardosh and Druckman (2006). This revision was made following the integration of the critical information from the Hannah-1 well that was not available at the time of the previous interpretation. The revised interpretation suggests that a significant part of the basin-fill is younger than was assumed.

The base Saqiye unconformity is recognized throughout the inland part of Israel. It is associated with an Oligocene uplift event that resulted in erosion and the development of an extensive drainage system and transport of sediments into the Levant Basin (see chapter 4.5). The incision and truncation on the Levant slope associated with the Red horizon conforms with the Early Oligocene paleogeography. The conformable, high-amplitude Red event within the basin is probably a condensed section at the top of the Upper Cretaceous to Eocene strata.

5.1.5 The Cyan horizon

The Cyan horizon is an unconformity surface that is characterized by a minor degree of truncation, incision and onlapping (Figs. 5.1, 5.3, 5.6). At the Yam West-1 and Yam Yafo-1 wells the marker is a low-amplitude event that is correlated to the top of the Oligocene-Lower Miocene Bet Guvrin Formation (Fig. 5.2, Table 3.1). The correlation is carried from the wells to the deep part of the basin where the Cyan horizon is identified as the top of several continuous,
Fig 5.9- Interpreted, NW-SE seismic profile from the Eratosthenes Seamount to the Levant margin. The profile shows a series of horst and graben structures, associated with Early Mesozoic rifting.
Reverse motion during the Tertiary have originated as a normal fault during the Early Jurassic and was reactivated in a northeastern part of the basin. The bounding fault of the Carmel-Caesarea block may have originated as a normal fault during the Early Jurassic and was reactivated in reverse motion during the Tertiary.
Fig 5.11- Interpreted SW-NE seismic profile showing the southern part of the Jonah Ridge. The ridge is composed of three units: (a) an Early Jurassic horst block, (b) a chaotic seismic package interpreted as a volcanic cone with highly dipping slopes, of Late Cretaceous to Early Tertiary age and, (c) a series of parallel, high-amplitude reflections interpreted as a carbonate buildup of Lower Miocene age.
high amplitude reflections below less reflective zones at a depth of about 4 to 5 sec (Figs. 5.5, 5.8, 5.9). In the deep basin the Cyan event is probably a condensed section at the top of the Oligocene to Lower Miocene strata.

5.1.6 The Purple horizon

The Purple horizon is the base of the Messinian evaporites. In wells at the eastern part of the basin it is correlated to a high-amplitude event near the top of the Ziqim Formation (Fig 5.2). Throughout the deep basin it is clearly recognized by the conspicuous change in seismic character from the underlying finely layered Middle-Upper Miocene strata to the overlying chaotic salt layer (Figs. 5.5, 5.8, 5.9, 5.10). The base Messinian unconformity, also known as the N horizon, is an Upper Miocene erosion surface associated with the huge drop of the Mediterranean sea-level and subsequent deposition of the Messinian evaporites (Hsu et al., 1973a,b; Finetti and Morelli, 1973; Ryan, 1978).

5.1.7 The Violet horizon

The Violet horizon is the top of the Messinian evaporites. In the deep basin the surface is clearly recognized by the conspicuous change in seismic character from the underlying chaotic salt layer to the finely layered Plio-Pleistocene strata (Figs. 5.5, 5.7, 5.8, 5.9). In the eastern part of the basin it is correlated to a high-amplitude event near the top of the Mavqiim Formation or to an erosion surface at the base of the Plio-Pleistocene section. In this area the Messinian evaporite layer is either very thin or completely missing and the Purple and Violet horizons merge (Figs. 5.3, 5.6). The top Messinian unconformity, also known as the M horizon, reflects a major lithofacies boundary associated with marine transgression and deposition of the hemipelagic Yafo Formation.

5.2 Time and Depth Maps

The seven seismic markers described above were interpreted on the seismic profiles of the A and B surveys throughout the study area (Fig. 1.3). A set of two-way-time maps was initially produced. These were converted to depth maps, layer by layer (layered cake method), through the multiplication of time and interval velocity grids (Figs. 5.12 to 5.19).

The reliability of time-to-depth conversion is commonly dependent on the quality of the velocity data. The main source of velocity information is check-shot surveys measured in wells. Well information in the study area is limited to the eastern part of the basin. Therefore, in a previous study Gardosh and Druckman (2006) applied a simplistic approach, using single interval velocity values.
Fig 5.13- Depth map of the Middle Jurassic, Blue horizon (meters below MSL).
Fig 5.14- Depth map of the Middle Cretaceous, Green horizon (meters below MSL).
Fig. 5.15- Depth map of the base Oligocene, Red horizon (meters below MSL).
Fig 5.16- Depth map of the top Lower Miocene, Cyan Horizon (meters below MSL).
Fig 5.17- Depth map of the base Messinian, Purple horizon (meters below MSL).
Fig 5.18- Depth map of the top Messinan, Violet Horizon (meters below MSL).
To improve the velocity control, in the present study we used two additional sources of velocity information: (a) stacking velocities that were extracted from the time processing of selected, regional lines from the A survey; and (b) horizon velocity analysis performed during the process of Pre-Stack Depth Migration of four lines. An interval velocity map was produced for each of the interpreted seismic horizons by integrating these velocities with the available check-shot data from the wells. Most of the interval velocity maps show east to west variations that are associated with increased depth of burial and degree of compaction from the margin into the basin. The resulting depth maps reflect these lateral velocity variations and are therefore considered more reliable than the previous versions of Gardosh and Druckman (2006). The depth maps were further controlled by the correlation of the various depth grids to the corresponding lithostratigraphic units in the 43 wells located within the study area (Fig. 1.3, Table 3.1).

6. Interpretation of Gravity and Magnetic Data

6.1 Gravity Maps

The raw gravity data measured during the acquisition of the B seismic survey were filtered and corrected for Eotvos, Free Air and Bouguer correction. The data were incorporated in the regional grid of the Levant area previously compiled by Rybakov et al. (1997). The combined Bouguer gravity map (Fig. 6.1a) shows a good fit with the new and old data sets. Negligible inconsistencies observed close to the tie line (white polygon) are explained by the poor gravity coverage of the B survey area.

The Bouguer gravity map (Fig. 6.1a) shows that a significant part of the area is occupied by high gravity values that reach up to 125 mgal. The composition of the crust underlying the Mediterranean is the subject of much discussion. Makris and Wang (1994) suggest an oceanic crust using seismic refraction, gravity and magnetic data, while Knipper and Sharaskin (1994) suggest that the crust is continental, based on their analysis of the tectonic history of the region. The high Bouguer gravity values are in agreement with the hypothesis that the crust underlying this part of the Levant Basin is dense. It is therefore either oceanic or intermediate in composition. Negative gravity values are typical for the continental crust of the African and Arabian plates.

Within the positive Bouguer gravity province a local gravity high is noticeable in the northern part of the basin (Fig. 6.1a). It should be noted that this gravity high was not highlighted in the previous gravity maps (Makris and Wang, 1994), probably due to poor gravity coverage in
Fig 6.1- Revised gravity anomaly maps of the Levant Basin. New gravity data set acquired in the area of Survey B (white polygon) is compiled with the gravity map of the Levant region (Rybakov et al., 1997): (a) Bouguer gravity, (b) Residual Bouguer gravity-second order polynomial. X-Y is the modeled section in Fig. 6.3.
Fig 6.2- Revised magnetic anomaly maps of the Levant Basin. New magnetic data set acquired in the area of Survey B (white polygon) is compiled with the magnetic map of the Levant region (Rybakov et al.,1997): (a) total magnetic intensity, (b) total magnetic intensity after reduction to pole. X-Y is the modeled section in Fig. 6.3.
this area. The positive province appears to be bounded to the SW and SE by two branches of relatively reduced gravity (zero Bouguer gravity close to the southern coastal plain of Israel) (Fig. 6.1a). The wide range of gravity values is probably associated with crustal changes; however the presence of the thick, light density sedimentary prism of the Nile River is an additional factor for the low gravity province in the south. The gravity steps and larger horizontal gradients in the southeastern part of the basin appear to represent a more complex pattern of near-surface density contrasts than in the southwestern area.

The effect of the shallower anomalies was enhanced by removing the regional deep trend up to the third order polynomial. The second order polynomial, shown in Figure 6.1b, is almost free of the deep crustal heterogeneities and shows the distribution of several high and low density blocks. The northern anomaly is further enhanced and is interpreted as a NE-SW trending, high density, shallow crustal block. It is limited to the SW by a gravity low, possibly associated with the effect of the light Nile prism. An additional more complex, but generally NE-SW trending, shallow crustal block is interpreted in the southern part of the basin. The narrow, elongated gravity low along the Israeli coast may be associated with the effects of the Plio-Pleistocene prism, or with deeper crustal heterogeneities that at this stage are not well resolved.

6.2 Magnetic Maps

The pattern of the magnetic anomalies (Fig. 6.2a) shows they are distributed randomly throughout the region, rather than associated with distinct, large-scale features as in the gravity anomaly map. We therefore assume that the magnetic anomalies are associated with local magmatic features and not with the deep crustal structures.

The low inclination of the Earth's magnetic field in the study area results in misplacement of the location and direction of the magnetic anomalies. An RTP (Reduction-To-Pole) calculation removes anomaly asymmetry caused by inclination and position of the anomalies above the causative magnetic bodies. It converts data which were recorded in the inclined Earth's magnetic field to what they would have been if the magnetic field was vertical. The RTP calculation was performed by the IGRF program (http://swdcd.b.kugi.kyoto-u.ac.jp/igrf/) assuming the midpoint of the area at 35E, 34N, the total intensity of the Earth's magnetic field is 44,763nT, its inclination is 49.6° and declination 2.7°. We also assume that the remanent magnetism is small compared to the induced magnetism.

The RTP map (Fig. 6.2b) preserves the overall magnetic pattern of the area; however it shows a shift of the positive and negative magnetic anomalies towards the NW. The accuracy of the RTP map is supported by the good correlation between the small, circular positive
anomaly found in the center of the area and a high density block of similar pattern observed on the residual gravity anomaly map (Fig. 6.1b).

The large positive magnetic anomaly found in the NW corner of the map (Fig. 6.2b) is the edge of the pronounced Eratosthenes magnetic anomaly (Ben-Avraham et al., 1976) located further to the NW. A large positive anomaly located in the SE part of the map (Fig. 6.2b) is associated with the Hebron magnetic anomaly (Rybakov et al., 1995) located inland further to the SE. Both anomalies are considered to be associated with Mesozoic volcanic bodies (Kempler, 1998; Rybakov et al., 1995).

An elongated NE trending positive magnetic anomaly intersects the NW trending Carmel magnetic anomaly near the coastline (Fig. 6.2b). The Carmel anomaly is associated with Jurassic basalts found in several wells near the coastline (Asher Volcanics) (Garfunkel, 1989; Gvirtzman et al., 1990). The NE trending anomaly in the center of the basin is possibly associated with a Jurassic or younger magnetic source.

6.3 Gravity and magnetic modeling (2D)

A regional geologic cross-section compiled from the seismic depth maps was used for modeling of the gravity and magnetic fields (Fig. 6.3). The section crosses the basin in a NW-SE direction more or less perpendicular to most of the anomalies observed in the gravity and magnetic maps (Figs. 6.1, 6.2). Calculation of the gravity and magnetic curves was carried out using Geosoft software.

Figure 6.3a shows the results of gravity modeling. The observed curve is taken from the Bouguer gravity map in Figure 6.1b. The modeled depth to the Moho is based on Ben-Avraham et al. (2002). The densities used for modeling are taken from well data. The basement density assumed in the central and western part of the section is 2.9 gr/mm³, corresponding to an oceanic or transitional type crust, whereas at the southeastern part it is assumed to be 2.75 gr/mm³, corresponding to a continental type crust (Ben-Avraham et al., 2002).

The good fit between the observed and calculated curves in the center of the basin supports the structural interpretation. A significant misfit is observed in the southeastern part of the section (Fig. 6.3a). The higher values of the calculated curve in this area can be explained in several ways: (a) the top of the crystalline basement is deeper than assumed; (b) the crystalline basement is either lighter; or (c) thicker than assumed (~22 km). The depth to the top of the crystalline basement in this area is relatively well constrained by the seismic data. Furthermore, in the Helez Deep-1 well located several tens of kilometers south of the edge of the section, the basement rocks were encountered at a depth of about 6 km, in agreement with the present structural interpretation. The density of the basement rocks in this well is 2.77 gr/mm³, similar
Fig 6.3- 2D Gravity and magnetic modeling of the Levant Basin (see Figs. 6.1 and 6.2 for location of the section): (a) observed and calculated Bouguer gravity, (b) observed and calculated magnetic intensity assuming continuous magnetic basement (light blue line), and (c) observed and calculated magnetic intensity assuming separate magnetic bodies (blue polygons).
to the value used for modeling. The most likely explanation for the misfit is therefore a thicker basement in the southeastern part of the section (Fig. 6.3a). This explanation is supported by a new deep refraction test (Netzeband et al., 2006), suggesting that the depth to the Moho near the coastline is 27 km, about 5 km more than previously assumed by Ben-Avraham et al. (2002).

A misfit between the observed and calculated curves is found also in the northern edge of the section (Fig. 6.3a). The higher calculated values may suggest a shallower basement than interpreted from the seismic data. However, it should be noted that the reliability of the observed gravity data in this area is reduced due to the smaller number of measurements.

Figures 6.3b and 6.3c show the results of the magnetic modeling. The observed curve is taken from the total magnetic anomaly map in Figure 6.2a. Two calculations were made. In the first one we assumed a magnetic basement extending along the entire basin (Fig. 6.3b). An optimal fit for this curve is reached when the top of the magnetic basement modeled is significantly shallower than the structural one (Fig. 6.3b). Additionally, the regional data suggests that the crystalline basement in the area is generally non-magnetic (Rybakov et al., 1999). Therefore, we consider the model in Figure 6.3b as unrealistic.

An alternative option is presented in Figure 6.3c. In this calculation we assume the presence of several highly magnetized bodies at a shallow depth within the sedimentary section. The southeastern body may correspond to the northern edge of the Hebron anomaly, which is assumed to be associated with Jurassic volcanics (Fig. 6.3c). A shallow magmatic body in this area may correspond to the Tertiary National Park Volcanics. A second magnetic body, located in the center of the basin is possibly associated with extrusive igneous rocks at the top of the Jonah Ridge (Fig. 6.3c). Several small bodies in the northwestern part of the section may correspond to the Eratosthenes high located nearby (Fig. 6.3c). It should be noted that the geometry and properties of these bodies as presented in Figure 6.3c are only tentative.

6.4. Discussion

The gravity and magnetic anomaly maps (Figs. 6.1, 6.2) and the modeled section (Fig. 6.3) shed new light on the deep structure of the Levant Basin. The high Bouguer gravity values in the central and northern part of the basin (Fig. 6.1a) indicate the presence of a dense crust, in the range of 2.9 gr/mm³. The residual map (Fig. 6.1b) suggests that the distribution of the positive gravity anomalies is related to several distinct blocks oriented NE-SW. These blocks are interpreted as fault-controlled basement highs (Fig. 6.3a) that were formed during Early Mesozoic rifting (Garfunkel and Derin, 1984; Gardosh and Druckman, 2006). Rifting was probably associated with extrusive volcanism as well as with intrusion of mafic rocks at a lower
crustal level. The combination of extension and magmatic intrusion resulted in the modification of the old continental crust and the formation of a thinned and heavy crust in the Levant Basin.

The NE-SW negative anomaly extending along the coastline is related to the transition from the intruded, heavy crust of the basin into the lighter (~2.75 gr/mm³), continental crust of the margin of the Arabian plate (Ben-Avraham et al., 2002). It is possible that the depth to the Moho at the eastern edge of the basin is deeper than previously assumed and is in the range of 27 km, as recently proposed by Netzeband et al. (2006).

The magnetic anomaly maps (Figs. 6.2a, b) and the modeled section (Fig. 6.3a, b) reveal the presence of several highly magnetized bodies. The distribution the magnetic anomalies is not directly related to the deep crustal structure of the Levant Basin indicated by the gravity anomaly maps. The modeled section suggests that these bodies are found within the sedimentary section and they are interpreted as extrusive volcanics. The Carmel anomaly at the northeastern edge of the basin is most likely associated with the Early Jurassic Asher Volcanics (Gvirtzman et al., 1990) found in the Atlit-I well and other boreholes in the Haifa area. An Early Jurassic age is also assumed for the Hebron anomaly (Rybakov et al., 1995) at the southeastern edge of the basin and the Eratosthenes anomaly at its northwestern edge.

The origin of the elongated, NE-SW oriented anomaly in the central part of the basin is less well understood. Its northeastern edge intersects the Carmel magmatic anomaly. Its southwestern edge is correlated to a positive gravimetric anomaly and a Mesozoic basement high, known as the Jonah structure (Folkman and Ben-Gai, 2004). The modeled section shows that the Jonah magnetic anomaly is found at a relatively shallow depth within rocks of Cretaceous to Tertiary age (Fig. 6.3c). It is speculated that this feature is related to Cretaceous and Tertiary eruptive episodes that are probably associated with a deeply rooted magmatic chamber.

7. Tectonic Evolution of the Levant Basin and Margin

7.1 Rifting Stage

7.1.1 Late Paleozoic to Early Mesozoic extensional structures inland

Rifting activity in the Levant region started in the Late Permian and continued in several pulses through the Triassic and Early to Middle Jurassic. Significant vertical movements and differential block motions took place during this period in northern Sinai, Israel and Syria (Fig. 7.1)(Goldberg and Friedman, 1974; Druckman, 1974, 1977; Freund et al., 1975; Druckman, 1984; Gelberman and Kemmis, 1987; Bruner, 1991; Druckman et al., 1995a; Garfunkel, 1998).
Fig 7.1- Neotethyan rifting stage- inferred Triassic to Early Jurassic horst and graben system of the Levant region. Inserted section shows the structural relations across the Helez fault, on the southern coastal plain (after Gardosh and Druckman, 2006).
Attesting to these motions are the thickness and facies changes of the Late Paleozoic and Mesozoic strata found in wells and outcrops in the region (for details see chapter 4).

The structural configuration revealed by this wide range of phenomena is an extensive graben and horst system extending east of the Mediterranean coastline (Fig. 7.1). Some of the faults that bound these structures are identified in seismic reflection lines (Gelbermann, 1995; Druckman et al., 1995b), while others are inferred from well data.

A major structural low extends in a NW-SE direction from northern Sinai to central Israel (the Hilal and Judea grabens)(Fig. 7.1). Towards the north this structure splits into two segments, the Asher graben trending to the NW and the southern edge of the Palmyra Trough trending to the NE. The Judea-Asher graben system is bounded on the west by three downstepping horst blocks: the Gevim high in the south and the Gaash and Maanit highs in the north (Fig. 7.1). The Helez fault is the eastern boundary of a structural low extending near the coastline west of the Gevim high (Fig. 7.1). Other, smaller scale structures are found in the southern Dead Sea area (Hemar and Massada highs) and probably also below the Syrian Arc structures of the northern Negev (Fig. 7.1)(Freund et al., 1975).

The timing of activity of this horst and graben system is broadly defined. Part of the system was active during the Permian (Garfunkel, 1998). Evidence for this initial stage in the inland Levant area is sparse due to limited well data. A significant tectonic phase took place during the Middle and Late Triassic (Anisian and Carnian) followed by a relatively quiet period during the latest Triassic (Garfunkel, 1998). A third and possibly the most extensive phase took place during the Early to Middle Jurassic (Liassic to Bathonian). The large amount of vertical offset during this phase is evident on the Helez fault (Fig. 7.1 insert) (Gardosh and Druckman, 2006), and abrupt thickness changes in the Liassic Ardon and Inmar formations (Goldberg and Friedman, 1974; Druckman 1977; Buchbinder and le Roux, 1992).

The Early Jurassic tectonic phase was followed by extensive extrusive magmatism in northern Israel (Asher Volcanics). Evidence for magmatic activity during this time, probably of more intrusive nature, is found also in various wells in central and southern Israel (Garfunkel, 1989; Rybakov et al., 1955). It is estimated that parts of the inland rift system were active at different rates during different times.

### 7.1.2 Late Paleozoic to Early Mesozoic extensional structures in the Levant Basin

The horst and graben system identified inland is observed on the seismic data of the Levant Basin offshore. The details of this system are more easily recognized in the central and western part of the basin than in the eastern part, where the sedimentary section was strongly affected by younger, contractional deformation (Figs. 5.8, 5.9).
Fig 7.2- Neotethyan rifting stage- Early Mesozoic extensional structures in the Levant Basin; (a) the northeastern part of the Jonah Ridge and (b) the southern part of the Yam High (see Fig. 7.1). Normal faults found east and west of the Yam High were reactivated as reverse faults.
The most prominent structures are two basement highs, the Jonah Ridge (Folkman and Ben-Gai, 2004) and Leviathan High (Figs. 5.8, 5.9, 7.1). A third structure, the Eratosthenes High, is partly covered by the seismic data and only its southeastern part is observed (Figs. 5.9, 7.1). The Jonah and Leviathan highs are structures 15 to 30 km wide and 80 to 100 km long trending in a NE-SW direction (Figs. 5.11, 5.12, 7.1). In the two structures the near-top-basement Brown horizon and the Middle Jurassic Blue horizon are markedly elevated from their surroundings. The identification of these horizons within the structures is somewhat tentative due to the correlation of seismic events across faults. However, the depth to the top of the basement level as derived from the seismic interpretation and depth maps is supported by the results of the gravity modeling presented in Chapter 6 (Fig. 6.3a). The Leviathan basement high is also observed as a large positive anomaly on the residual Bouguer gravity map (Fig. 6.1).

The two structures are bounded by sets of high-angle normal faults with vertical offsets in the range of several hundred meters to a few kilometers. The structural lows on the two sides of the structures are characterized by dipping events, forming asymmetric grabens (Figs. 5.9, 7.2a). The grabens are filled with 4-8 km thick sections of Permian to Middle Jurassic strata.

Several smaller basement highs, about 10 km wide and 20 km long trending in a NE-SW direction are identified south and north of the Jonah and Leviathan ridges (Figs. 5.12, 5.13, 7.1). A larger structure is partly revealed by the seismic data in the southeastern part of the basin. The Yam High is identified by a series of deep-seated normal faults found west of the Yam-2 and Bravo-1 wells (Figs. 7.1, 7.2b). The near-top-basement level is downfaulted to the west, forming a set of down-to-the-basin fault blocks. The thickness of the Brown to Blue interval shows a marked increase from about 1 km in the east to about 3 to 4 km in the west (Fig. 7.2b). The Yam High probably extends several tens of kilometers to the NE (Fig. 7.1). Its eastern boundary is marked by a series of faults located near the Yam-2 and Bravo-1 wells (Fig. 7.2b). These are interpreted as normal faults that were later reactivated in a reverse motion.

Unlike the situation inland, the timing of activity of the horst and graben system within the basin is only broadly defined due to the uncertainty associated with the age of the seismic markers. The asymmetric grabens near the Jonah and Leviathan ridges suggest syntectonic deposition through the Late Paleozoic to Middle Jurassic periods. Some of the faults bounding these structures appear to have been active also during post Middle Jurassic time (Fig. 7.2a), however, it is assumed that this late activity was only minor and during the Late Jurassic the structures were well developed and formed submarine topographic highs. The faults on the western edge of the Yam High do not offset the Middle Jurassic horizon, thus suggesting a Late Paleozoic to Triassic age of this structure (Fig. 7.2b).
The near-top-basement to Middle Jurassic interval shows significant thickness variations across the basin. It is about 3 km thick near the Mediterranean coastline and the Eratosthenes and Leviathan highs and about 5 to 8 km thick in the central part of the basin (Figs. 5.8 to 5.10). The middle part of the basin termed here the Central Levant Rift (Fig. 7.1) is interpreted as the main depocenter associated with the Late Paleozoic to Middle Jurassic rifting activity. Intra-rift highs such as the Jonah Ridge were formed within this large depocenter. Localized lows may have been partly filled with a thick volcanic series (Asher Volcanics), as suggested by Gardosh and Druckman (2006).

7.1.3 Neotethyan rifting in the Levant area

The onshore and offshore horst and graben system was formed under an extensional regime associated with major plate motions and the opening of the Neotethys Ocean in the Levant area and north of it (Garfunkel and Derin, 1984; Garfunkel, 1998). The Levant rift system was bounded by the Arabian Massif to the southeast and the Eratosthenes continental block to the west and thus was probably separated from the main body of the Neotethys Ocean that extended north of the Eratosthenes Seamount and was later consumed underneath Cyprus and southern Turkey (Garfunkel, 1998, 2004; Robertson, 1998). Four main rifting episodes are recognized: during the Permian; early Middle Triassic; Late Triassic; and Early to Middle Jurassic. The amount of extension during each of these phases is not well established.

The general strike of the normal faulting in the Levant Basin is NE-SW; accordingly, we interpret an extension in a NW-SE direction perpendicular to the strike of the faulting. The extension may have been accommodated by NW-SE trending transform faults within the basin (Fig. 7.1). Garfunkel and Derin (1984) and Garfunkel (1998) postulated a similar strike-slip fault along northern Sinai to explain the separation and northward motion of the Tauride and Eratosthenes blocks. An apparent lateral offset in the northern edge of the Jonah ridge and the southern edge of the Leviathan High (Fig. 7.1) may be explained by strike–slip faulting.

The structural scheme presented above for the Neotethyan rifting is not in accord with extension and spreading of the eastern Mediterranean in an N-S direction accompanied by a transform fault along the eastern Mediterranean shoreline, as proposed by Dewey et al. (1973), Bein and Gvirtzman (1977), Robertson and Dixon (1984) and Stampfli et al. (2001).

An important question regarding the nature of the Neotethyan rifting processes is whether emplacement of new oceanic crust took place in the Levant area. Apart from faulting no pronounced disruption or conspicuous lateral variations in the seismic properties of either the upper part of the basement or the Late Paleozoic to Middle Jurassic interval are observed in the seismic data set. The basement layer in the central part of the basin does not show any
characteristics of oceanic crust such as described in other passive continental margins (Klitgord and Hutchinson, 1988) or in the central part of the Mediterranean Sea (Finetti, 1985; Avedik et al., 1995). The normal faults and basement highs associated with the rifting stage are well preserved. Although a considerable amount of Triassic and Jurassic volcanics may exist within the basin, their equivalent units in the onshore area (Asher Volcanics) do not show MORB characteristics associated with sea-floor spreading. Thus, the occurrence of spreading and introduction of new oceanic crust cannot be supported by the present data.

In view of the above discussion, it is suggested that the Early Mesozoic rifting that started on the northern edge of Gondwana may not have reached the spreading phase in the area of the present Levant Basin (Gardosh and Druckman, 2006). A recent analog for this scenario is the northern part of the Red Sea. There, the rift is continental with only a nucleation of an oceanic spreading center and an early magmatic phase. An oceanic spreading center has developed only in the southern and central parts of the Red Sea (Martinez and Cochran 1988; Bohannon and Eittreim 1990; Cochran, 2001).

The thinned crust in the center of the Levant Basin (Fig. 2.1) may be explained in two ways: (a) extension and stretching on basement-involved sets of normal faults; (b) magmatic underplating and thermal erosion at the base of the crust (Hirsch et al., 1995). It should be noted that the amount of extension on the Early Mesozoic faults appears to be limited as most of them are high-angle normal faults and lack the listric character that is often observed in other rifted continental margins. The high-density of the thinned crust that is interpreted from seismic velocities (Fig. 2.1) (Ben-Avraham et al., 2002) and is indicated by the modeling of gravity data in the present study (Fig. 6.3c) may be associated with intrusion of mantle material into the old Pan-African continental crust of northern Gondwana (Hirsch et al., 1995).

7.2 Post-Rift, Passive Margin Stage

Rifting and normal faulting activity in the Levant by and large ceased during Middle to Late Jurassic (Garfunkel, 1998). A post-rift, passive margin stage was initiated at this time as a result of crustal cooling and thermal subsidence (Garfunkel and Derin, 1984; Garfunkel, 1989; ten Brink, 1987). A faster rate of subsidence in the central part of the basin compared to that on its margins may have initiated the formation of the 'Mesozoic depositional Hinge-Belt' along the Mediterranean coast.

The Hinge-Belt is characterized by a pronounced facies change in the Mesozoic section, from shallow water carbonates and sandstones in the east to fine-grained, pelagic and hemipelagic carbonates and siliciclastics in the west (Derin, 1974; Bein and Gvirtzman, 1977; Flexer et al., 1986). This pronounced facies change, which was detected in many onshore and offshore
Fig 7.3- Passive margin stage- (b) isopach map of the Middle Jurassic to Middle Cretaceous interval, and (a) geologic section through the Levant margin showing the two dominant depositional styles: backstepping and aggrading, shallow marine carbonate platforms on the east; and deepwater, siliciclastic and carbonate turbidite systems on the west (from Gardosh, 2002). Blue and green lines denote seismic horizons.
wells (Fig. 7.3), indicates the development of a deep marine basin in the Levant area, bordered by a slope and a shallow-marine shelf.

There is some ambiguity regarding the time of initiation of this passive margin-type, depositional profile. Our stratigraphic analysis shows that all the Triassic as well as the first four Jurassic depositional cycles (Pliensbachian to Bathonian) are generally dominated by shallow-marine-type rocks. However, the allochthonous Bathonian section in the offshore Yam Yafo-1 well (Fig. 4.2) contains mass flows that were obviously deposited in a deeper marine environment (see chapter 4.3). It is therefore assumed that the allochthonous succession in the Yam Yafo-1 well indicates an initial stage of basin evolution within a predominantly platform setting.

A deep marine depositional environment was well established throughout the Levant basin since Oxfordian to Kimmeridgian times. The corresponding rock interval in the offshore Yam Yafo-1, Yam-2 and Yam West-1 well (Delta Formation) is composed of carbonate mud of deep marine origin with transported material (Fig. 7.3) (Derin et al., 1990; Druckman et al., 1994, Gardosh, 2002).

The deepening of the basin and the change in depositional environments is further indicated by the change in seismic character delineated by the Blue horizon. A relatively continuous high-amplitude reflection series, interpreted as shallow-marine strata below the Blue horizon changes upward to a discontinuous, low-amplitude, occasionally shingled seismic facies (between the Blue and Green horizons). The latter is interpreted as fine-grained strata of deep-marine origin with mass transported components (Fig. 5.1).

The strata architecture along the passive margin was significantly affected by eustatic changes. Gradual sea-level rise resulted in backstepping and aggradation of the carbonate platforms, whereas fast sea-level rise resulted in drowning of the carbonate platforms. Sea-level drops were associated with bypass of the shelf and deposition of siliciclastic and carbonate gravity flows on the slope and in the basin. These are remarkably demonstrated by the siliciclastic, turbidite system of the first two Cretaceous depositional cycles (Berriasian to Hauterivian and Hauterivian to Lower Aptian; Fig. 4.3).

The passive margin succession reaches a thickness of 2500 to maximal 3500 m (Fig. 7.3a). Its depocenter extends from the present coastline area to about 70 km westward. In the central part of the basin and near the Eratosthenes High the thickness of this succession is reduced to several hundreds up to 2000 m (Fig. 7.3a). In this distal part of the basin the rate of accumulation of deepwater sediments was significantly slower than near the margin.

Thick accumulations of the passive margin stage (between the Blue and Green horizons) are located adjacent to the Early Mesozoic horsts (east and west of the Jonah Ridge in Fig. 7.2a).
It is assumed that these areas continued to act as localized depocenters, accumulating clastic material that was shed from the nearby submarine topographic highs. The tops of the Jonah, Leviathan and Eratosthenes structures may have reached shallow water level and hosted isolated carbonate buildups of Middle-Late Jurassic and Middle Cretaceous age.

### 7.3 Convergence Stage

#### 7.3.1 Late Cretaceous to Tertiary contractional structures inland

The termination of post-rift, passive margin stage coincides with the onset of a regional contractional deformation phase associated with the formation of the 'Syrian Arc' (Krenkel, 1924) or the 'Levantid' (Picard, 1959) fold belt. This structural element is an 'S' shape mountain belt extending from the Palmyra Mountains in Syria to the Lebanon and Anti-Lebanon Mountains; through the Judea Mountains and Negev anticlines and into the northern Sinai anticlines (Fig. 7.4) (Picard, 1943, 1959; Ball and Ball, 1953; Bentor and Vroman, 1954, 1960; de Sitter, 1962; Gvirtzman, 1970; Bartov, 1974; Neev et al., 1976; Horowitz, 1979; Eyal and Reches, 1983; Lovelock, 1984; Beydoun, 1988; McBride et al., 1990). The fold belt extends further to the southwest below the thick Mio-Pliocene sediments of the Nile Delta (Aal et al., 2000) and into the Western Desert of Egypt (Said, 1990).

The Syrian Arc belt consists of a series of surface and subsurface anticlines that strike in an ENE, NE and NNE direction. Most of the structures are open flexures and folds, in some areas forming broad anticlinorial upwarps. Several characteristic geometries are identified: the folds of the central and northern Negev are all strongly asymmetric with steep flanks on the southeast, while the Hebron anticline is strongly asymmetric on the west (Fig. 7.4). Other structures, such as the Fari'a anticline (Fig. 7.4), are roughly symmetric box folds with steeply dipping flanks on both sides (Flexer et al., 2005).

The Syrian Arc folds are frequently associated with reverse faulting. The concept of reverse faults at depth, as first suggested by de Sitter (1962), was substantiated by several deep oil wells in southern and central Israel in which strata repetitions were discovered. Freund et al. (1975) suggested that these reverse faults originated as normal faults during the Early Mesozoic rifting stage. Compelling evidence for inversion of the older extensional structures has been supplied from various parts of the Syrian Arc fold belt in onshore Israel and Syria (Davis, 1982; Gelbermann and Kemmis, 1987; Bruner, 1991; Best et al., 1993, Chaimov et al., 1993; Druckman et al., 1995a).

The timing of the Syrian Arc deformation inland is debated, but the main range of ages given by various authors is Turonian to Neogene (Picard, 1943; Bentor and Vroman, 1954, 1960; Freund, 1965; Bartov, 1974; Eyal and Reches, 1983). Walley (1998), who summarized
Fig 7.4 – Convergence stage-Syrian Arc fold structures of the Levant region. Combined depth map of the Middle Cretaceous level: onshore after Fleischer and Carlson (2003) top Judea map; and offshore depth map of the Green horizon in present study. Note Syrian Arc I and II folding phases. The Syrian Arc II structures offshore are projected from the top Lower Miocene level.
Fig 7.5- Convergence stage- contractional structures in the Levant Basin: (a) profile from the southeastern part of the basin showing two folding phases, 1= Syrian Arc I (Senonian) and 2= Syrian Arc II (Oligo-Miocene); (b) profile from the southeastern part of the basin showing Syrian Arc I+II folds, note the onlapping of the Oligo-Miocene strata on the tilted, faulted and folded block apparently deformed during Syrian Arc II folding phase; (c) profile from the eastern part of the basin showing Syrian Arc I+II folds; (d) profile from the northeastern part of the basin showing Syrian Arc II folds
observations on the Syrian Arc system from Lebanon and other parts of the Levant, suggested two main episodes of deformation: Syrian Arc I during Coniacian to Santonian times; and Syrian Arc II during Late Eocene to Late Oligocene times. Mimran (1984) found a ‘post Neogene’ age for the latest folding phase in the Fari’a anticline (Fig. 7.4), while Freund (1965) infers a post Pliocene continuation of folding in some structures. Eyal (1996), who studied the properties of Syrian Arc folds and related structures throughout Israel and Sinai, suggested that the Syrian Arc stress field in the Levant area may have persisted into the Miocene and possibly to the present.

7.3.2 Late Cretaceous to Tertiary contractional structures in the Levant Basin

Syrian Arc type contractional deformation is observed in the offshore seismic data throughout the Levant Basin and margin. Two folding episodes are identified, and following Walley (1998, 2001), these are termed here Syrian Arc I and II (Fig. 7.5a).

Syrian Arc I structures are predominant in the eastern part of the basin some 50 to 70 km west of the coastline (Fig. 7.4). The dominant deformational style in this area is of high-amplitude and short wave length anticlines and synclines accompanied by high-angle thrust faults (Figs. 5.3, 7.5a). The deformation affects the Middle Cretaceous unconformity (Green horizon) and older strata; therefore a Senonian age for this folding is inferred (Figs. 5.3, 5.6, 7.5a).

The time and style of deformation of the Syrian Arc I structures offshore appear to be similar to the folds of the northern Negev inland. The size of the folds ranges from 10 to 30 km in length and 5 to 10 km in width. Their height ranges from several hundred to more than 1000 m and their flanks dip 100-300. Many of the folds are asymmetric with steep flanks on the east or southeast (Fig. 7.5a, b).

The thrust faults dip 65-75° and are traced down into the basement. Most of the faults offset up to the Middle Cretaceous unconformity (Green horizon), although rarely, they also penetrate the overlying Tertiary strata (Figs. 5.3, 5.6, 7.5a, b). In some areas evidence for inversion of the older, extensional structures is observed. For example, the reverse fault west of the Yam West-1 well may have been a normal fault associated with down-to-the-basin, Early Mesozoic faulting (Figs. 5.3, 7.2b). Many of the faults resemble positive flower structures and contain supporting limbs; thus a certain amount of wrenching or transpression is inferred.

Syrian Arc II contractional deformation in the offshore is more complex and contains two elements. The first element is a series of low-amplitude long wave folds that are in some places superimposed on the older Syrian Arc I structures (Figs. 7.5a, c). The deformation affects the Lower Miocene unconformity (Cyan horizon) and older strata; therefore a Middle Miocene age of this folding phase can be inferred. Syrian Arc II folds are found throughout the basin and also
in its deep part east of the Eratosthenes high (Fig. 7.4). Reverse faulting and inversion are generally not associated with this style of deformation; however, some of the structures in the deep part of the basin are located above deep-seated basement highs, possibly suggesting some reactivation of the Early Mesozoic horst and graben system (Fig. 5.9). The time span of the Syrian Arc II deformation phase is not well constrained. Folding could have started in the Late Eocene, as proposed by Walley (1989, 2001), and could have continued through the Early to Middle Miocene. The base Messinian unconformity (Purple horizon) is generally not affected or very mildly deformed (Figs. 7.5a-d). Therefore an upper age limit for the Syrian Arc II deformation in the Levant Basin is the latest Miocene.

A second element associated with the Syrian Arc II deformation is uplifting and tilting. Figure 7.5b shows an uplifted block at the western end of the Syrian Arc I fold belt (Fig. 7.4). The Middle Cretaceous unconformity (Green horizon) is tilted westward and the uppermost part of the structure is elevated some 3000 m from the basin floor (western part of the profile in Fig. 7.5b). The Oligocene to Lower Miocene section onlaps the elevated structure on the east; thus indicating filling of a negative relief (Fig. 7.5b). Uplifting and onlapping of Neogene strata characterize the entire western edge of the Syrian Arc I fold belt in the eastern part of the basin (Fig. 7.4 and seismic examples in Figs. 5.3, 5.6, 5.8.). The age of this event is estimated as Oligocene to Upper Miocene (pre-Messinian).

At the northeastern part of the basin a younger deformation is observed. In the Carmel-Cesarea area the Mesozoic and Cenozoic section is uplifted near the coastline and the thin Plio-Pleistocene section on top of the structure thickens significantly west of it (Fig. 5.10). It is therefore suggested that in this area the uplifting continued through the Plio-Pleistocene.

7.3.3 Neotethyan convergence in the Levant area

It is widely accepted that the Syrian Arc contractional deformation of the Levant area is associated with the collision of the African-Arabian and Eurasian plates (Reches and Eyal, 1983; Garfunkel, 1998; Walley, 1998, 2001; Flexer et al., 2005). A late Early Cretaceous convergence initiated a northward-dipping subduction zone within the southerly Neotethys oceanic basin (Robertson, 1998) that eventually progressed to collision and accretion in the present-day area of Cyprus and southern Turkey. This activity is manifested at a lower intensity by the Syrian Arc contractional structures of the Levant.

Reches and Eyal (1983) and Eyal (1996) described the Syrian Arc stress field (SAS) associated with this deformation. According to these authors the SAS regime has been active from the Turonian to the present. This suggestion is generally supported by the data from the Levant Basin. The long-term contraction is marked by phases of more intense deformation.
The Senonian Syrian Arc I phase is characterized by localized folding, partly controlled by the location of the deep-seated, Early Mesozoic normal faults. Walley (1998) proposed that this phase reflects an ocean-ocean collision in the southern Neotethys. Thickness and facies variations of the Senonian strata in the Levant area are relatively limited, therefore it is speculated that no significant relief was formed between the Levant margin on the east and the deep basin on the west during the Syrian Arc I deformation.

The Syrian Arc II phase that started in the Paleogene is possibly related to the progression of convergence and to a continent-continent collision on the southern Neotethys margin (Walley, 1998, 2001). The wide extent of this deformation that was previously identified only in several locations inland is revealed in the offshore data (Fig. 7.4). It is assumed that many of the older, Syrian Arc I folds onshore and offshore were reactivated while new structures were continuously being formed. The offshore seismic data further indicates that the Syrian Arc II contractional deformation continued through the Miocene (Eyal, 1996) and did not end in the Late Oligocene as proposed by Walley (1998, 2001).

An important element of the Syrian Arc II deformation shown by the offshore seismic data is uplifting and tilting on the Levant margin (Fig. 7.5b). Walley (1998) proposed that the main uplifts of the Lebanon and the Anti-Lebanon mountains occurred during this phase. It is assumed that uplifting of the mountainous backbone of Israel also took place during the Oligo-Miocene and resulted in erosion, westward transport of clastic sediments and huge accumulations within the deep part of the Levant Basin. Uplifting continued locally in the Carmel-Caesarea area during the Plio-Pleistocene, possibly in conjunction with the activity on the nearby segment of the Dead Sea Transform (Yagur fault) (Ben-Gai and Ben-Avraham, 1995).

8. Erosion Processes and Mass Transport on the Levant Slope

8.1 Rifting and Passive Margin Stages

Erosion and transport of sediments from elevated areas in the south and east to the basin in the west and northwest have been taking place throughout the Mesozoic and Cenozoic history of the region. The earliest evidence for coarse, clastic sediment accumulation is found in the Helez Deep-1 well in the southern coastal plain of Israel (Fig. 1.3). The Erez Conglomerate (Druckman, 1984) is a few hundred meters thick, built of polimictic carbonate breccia of lower Late Triassic age (Fig. 4.1). Druckman (1984) described this unit as a fault breccia that was eroded from a nearby horst block (Gevim High, Fig. 7.1) and accumulated on the downthrown side of the Helez fault. The Triassic to Middle Jurassic horst and graben morphology could have
accommodated a well-developed drainage system, transporting fine and coarse siliciclastics. During lowstands, e.g., between the Jurassic cycles 3 and 4 (Fig. 4.2), the Inmar sands may have been transported from the distal Sinai and Negev hinterland toward the basin in the north and west.

During the passive margin stage deepwater turbidities accumulated on the slope and within the basin, west of the shallow-marine shelf (Fig. 7.3) (Gardosh, 2002). Thin oolitic and micro-conglomerate beds found in the Upper Jurassic cycle offshore (Yam-2, Yam West-1 and Yam Yafo-1 wells; Derin et al., 1990; Druckman et al., 1994) are the products of submarine, mass transport processes. The distribution and paleogeographic extent of these units are not fully resolved.

The Lower Cretaceous Gevar'am Formation reflects an extensive deepwater turbidite system that developed on the Levant slope during a prolonged lowstand (Fig. 4.3). This unit comprises a thick series of hemi-pelagic dark shale found in various onshore and offshore wells. In the southern coastal plain it fills a 1 km deep canyon, incised within the Jurassic shelf (Cohen, 1976). The Gevar'am Canyon was the main conduit for submarine turbidite flows that transported significant amounts of quartz-toze sands into the basin from the elevated area on the southeast. Several sand beds up to 20 m in thickness, found within the Gevar'am shale offshore (Yam-2 and Yam West-1 wells, Fig. 7.3), are interpreted as distal slope fans and questionably basin floor fans that are associated with the Gevar'am Canyon and channel system (Gardosh, 2002). Lower Cretaceous deepwater slope and basin floor fans are likely to be found in other parts of the southern and central Levant Basin (e.g., Mango-1 well, Fig. 1.2).

The Middle Cretaceous section of the Levant margin and basin, although dominated by carbonate strata, show evidence of clastic sediment transport. The Aptian-Albian slope deposits of the Talme Yafe Formation contain coarse-grained carbonate breccias found in various wells in the coastal plain (Cohen, 1971; Bein and Weiler, 1976). These were probably eroded from the shelf and deposited by gravity flows on the slope during short term Aptian–Albian lowstands (Gardosh, 2002).

Another Middle Cretaceous erosional feature of the Levant slope is the Item Canyon. The Item Formation is a 2000 m thick carbonate series of Cenomanian age (Fig. 4.3), found in the Item-1 well (Fig. 1.3). The origin of this unit was previously not fully understood. The new offshore seismic data provides evidence for erosion at the base of the unit, indicating deposition within a deep submarine slope canyon (Fig. 4.6). The geometry of the Item Canyon is at present not well resolved because of the later Syrian Arc deformation.

A pronounced middle Turonian lowstand associated with karstic phenomena and quartz-toze sandstone deposition is well documented in the shallow-marine platform from the
Fig 8.1 - Convergence stage-the Tertiary fill of the Levant Basin: (a) Stratigraphic section through the Levant Basin and Margin, and (b) isopach map of the Oligocene to Upper Miocene interval. The great thickness of the Tertiary section in the basin (more than 3 km) resulted from tectonic uplift, eustatic sea-level falls and intense erosion and mass transport across the Levant slope. See location of the section on the isopach map.
Long distance transport of Nubian sandstone

Long distance transport of Nubian, Hazeva and Hordos sandstone
Fig 8.2- The Tertiary, submarine canyon and channel system of the Levant margin, shown on the depth maps of: (a) base Oligocene, (b) top Lower Miocene (c) base Messinian and (d) top Messinian. The main directions of siliciclastic sediment supply are marked with black arrows.
Negev to the Galilee area (Fig. 4.3) (Sandler, 1996). It is possible that Turonian sands bypassed the shelf and were transported westward by submarine turbidite flows to the slope and basin areas, although these rocks were not encountered yet in any offshore wells.

8.2 Convergence stage

8.2.1 Submarine canyon morphology

Extensive erosion and incision on the Levant slope, followed by the accumulation of several kilometers thick Tertiary fill within the basin took place during the convergence stage (Fig. 8.1). These phenomena reflect the combined effects of tectonic uplifting and tilting of the Levant margin and the mountainous backbone of Israel, the uplifting of the Arabian Shield and eustatic sea-level drops (Fig. 4.4). A highly developed submarine canyon and channel system evolved during the Early Oligocene and persisted till the Late Pliocene (Figs. 8.2, 8.3). The system included several main conduits generally oriented in NE-SW and E-W directions. These are, from south to north: the Afiq, Ashdod, Hadera, Caeserea, Atlit and Qishon canyons (Fig. 8.2). Some of these erosional features were previously identified in onshore wells (the Afiq and Ashdod canyons, Gvirtzman and Buchbinder, 1978; Druckman et al., 1995b; Buchbinder and Zilberman, 1997). The offshore seismic data reveal the extent of the Tertiary drainage system along the entire eastern part of the Levant Basin up to 70 km west of the present-day coastline. The system is built of superimposed canyon incisions revealed by the topography of the base Oligocene, Lower Miocene, base Messinian and top Messinian unconformities (Figs. 8.2a-d).

The major canyons are more or less vertically stacked and are straight to slightly curvilinear. Secondary, subsequent channels confined by the morphology of pre-existing fold structures flowed into the main canyons. The Afiq and Ashdod canyons are deeply incised. The depth of the Afiq Canyon is up to 1500 m onshore (Druckman et al., 1995b) and several hundred meters offshore (Fig. 8.3b). The Ashdod Canyon cuts a 700 m-deep gorge into Mesozoic strata some 40 km offshore (Fig 8.3a). The incisions in the northern channels (Hadera, Caesarea, Atlit and Qishon) are limited to the upper slope, near the present-day coastline.

8.2.2 Oligocene to Lower Miocene canyon system

The Oligocene tectonic activity and sea-level falls resulted in considerable shedding of clastic material into the Levant Basin. Denudation was extensive, exposing Nubian sandstone terrain in far proximal areas. Widespread transport of clastic material initiated a system of submarine canyons, depicted by the Red seismic horizon (Fig. 8.2a). Oligocene continental deposits are absent in Israel. Apparently, clastic material was not stored inland but was directly transported to the basin.
Approximately 250 m of Oligocene coarse-grained clastics (sandstones and conglomerates, named "Ashdod Clastics") were penetrated in the Ashdod wells, located in the southern coastal plain where the Ashdod Canyon meets the present-day shoreline (Fig. 8.2a) (Buchbinder and Siman-Tov, 2000; Buchbinder et al., 2005). The age of the Ashdod Clastics is not strictly defined. It may range from upper Rupelian to Lower Chattian (zones P19 to P21 in Fig. 4.4; Buchbinder et al., 2005). Core #2 from the Ashdod-2 well reveals a conglomerate consisting of boulders, imbricated pebbles (mostly of Cretaceous limestones and dolostones) and bioclastic sandstones of turbidite origin. Two intervals of highly porous sandstones, 20 m to 50 m thick, were detected in the Ashdod Clastics section. Their Spontaneous Potential (SP) and Gamma-Ray (GR) log response exhibit blocky to fining-upward patterns. Other parts of the section show a serrated log motif, reflecting interbedded conglomerates, sandstones and marls (Buchbinder et al., 2005). The clastic succession is interpreted as canyon fill or proximal slope fan, deposited within the submarine, Ashdod Canyon that was formed during the lower Oligocene (Figs. 4.4, 8.2a).

Although no significant sandy deposits were encountered in onshore wells along the Afiq Canyon (Fig 8.2a), Martinotti, (1981) reported a Cretaceous boulder within hemi-pelagic middle Oligocene sediments in the Gaza-1 well. Apparently, sands either bypassed the present-day, onshore part of the Afiq canyon and were deposited offshore, or were eroded during the repeated canyon entrenchments in the Miocene.

According to new findings from the Qishon-Yam-1 well (Fig. 1.2) (Schattner et al. 2006), the Qishon-Yizrael valley in the north part of the country was formed already in Oligocene times. Therefore, clastic sediments from the Damascus- Ein Gev Basins could have been transported through the Qishon and Atlit Canyons to the Levant Basin (Fig 8.2a) (see below).

In the offshore area, high-amplitude, discontinuous seismic events that were identified within Oligocene canyons may be interpreted as stacked, migrating channel systems (Fig 8.3a,c). This seismic pattern probably indicates transported, coarse clastic material. It is highly probable that clastic material was further transported into the basin and was deposited as sandy basin-floor fans. On the seismic data these may correspond to high-amplitude seismic events, which were identified above the Red horizon near the distal outlets of the main canyons (Fig. 8.3d).

The Lower Miocene is characterized by a prolonged lowstand (~6 my) indicated by the absence of marine Burdigalian sediments inland (Fig. 4.4). This lowstand occurred during a period of global eustatic rise of sea-level (Haq et al., 1977; 1978, Fig. 4.4) and may thus indicate a tectonic uplift (Syrian Arc II phase). The lowstand was accompanied by intensive erosion and transportation of clastic material into the basin. A Lower Miocene sand interval, 17 m thick, was
Fig 8.3- Tertiary submarine incision and deposition of clastic sediments in the Levant Basin: (a) the incised Ashdod Canyon; (b) the incised Afiq Canyon; (c) high-amplitude, discontinuous seismic reflections within the Hadera channel (1), interpreted as stacked, migrating channel systems; and (d) high-amplitude seismic reflections near the mouth of the Ashdod Canyon(2), interpreted as basin-floor fan lobes.
encountered in the Gaza-1 well, in the southern coastal plain (Figs.1.3, 4.4). It may represent the proximal tail of much more extensive accumulation in a distal basin position. The seismic data shows high-amplitude events in the upper part of the Oligocene to top Lower Miocene interval, below the Cyan horizon (Figs. 8.3b,d). These seismic events may correspond to clastic accumulations in a distal basin position.

Notably, continental clastic sediments of the Hazeva and Hordos formations started to accumulate in proximal inland areas in Early Miocene times. The Hordos sedimentation in the Galilee initiated before 17 my (Shaliv, 1991) and the Hazeva Group in the Negev started to accumulate before 20 my (Calvo and Bartov, 2001). The Hazeva continental deposition was occasionally interrupted by Syrian Arc or other tectonic movements which may correspond to the significant erosional truncation event of the Zefa Member (Calvo and Bartov, 2001). This event could have resulted in a long distance transportation of clastic material from southern Israel into the basin through the Afiq and Asdod Canyons (Fig. 8.2a,b).

In the north, the Nukev, En Gev and Hordos sands (of the Golan area) could have also been remobilized and transported through the Qishon Valley to the Levant basin. According to Shaliv (1991) the similarities between the En-Gev sands and the continental sediments of the Damascus basin suggest that the two sites are, in fact, parts of the same basin. This indicates the existence of extensive source for siliciclastics in the north (Fig. 8.2a,b).

In summary, the Oligocene to Lower Miocene period is characterized by widespread transport in submarine canyon and channel system along the entire Levant margin. The provenance of the clastic material was long distance sites of Nubian sandstone (which was already exposed in Sinai, Negev, and Trans-Jordan) during the early Oligocene phase, and both Nubian sandstone and remobilized Hazeva and Hordos sands during early Miocene times (Fig. 8.2a,b).

### 8.2.3 Middle to Upper Miocene canyon system

The Serravallian to early Tortonian time is characterized by a long term lowstand (Serravallian crisis). This lowstand is associated with the Bet Nir Conglomerate found in outcrops inland; and probably also with 250 m thick conglomerate interval found in the Nahal Oz-1 well, located in a proximal part of the Afiq Canyon, at the southern coastal plain (Fig. 8.2b). Druckman et al. (1995b) described in this part of the canyon erosion and slumping event of Langhian to Serravallian age (2nd erosion; Druckman et al., 1995, figure 4). Further updip, the Beer Sheva Canyon was excavated at this time and later filled by Late Miocene sediments of the Pattish cycle (Fig. 4.4) (Buchbinder and Zilberman, 1997).
The Serravallian lowstand is associated with submarine incision on the Levant slope, depicted by the Cyan seismic horizon (Fig. 8.2b). The canyon pattern closely followed the older, base Oligocene system although the degree of incision was reduced (Fig. 8.3a,c). Coarse clastic material was transported basinward through the Middle to Upper Miocene canyon system (Fig. 8.2b). Discontinuous, high-amplitude seismic events found above the Cyan horizon (Fig. 8.3b) may be associated with channel-fill deposits and basin floor fans. Evidence from several wells recently drilled in the Afiq Canyon offshore support this interpretation. The provenance of the siliciclastic material was long distance sites of Nubian sandstone and remobilized Hazeva and Hordos sands (Fig. 8.2b).

8.2.4 The Base-Messinain-evaporite canyon system

Substantial subaerial and submarine denudation of the North African and Levant continental margin took place during the Messinian salinity crisis and lowstand (Fig. 4.4) (Ryan, 1978; Gvirtzman and Buchbinder, 1978). On the Levant slope the base Messinain canyon system is depicted by the Purple seismic horizon (Fig. 8.2c). The degree of incision at the base of the evaporite section is smaller than previous erosion phases (Fig. 8.3a,c) possibly due to reduced runoff during this arid period.

The mountainous backbone of Israel was well developed at this time forming a topographic barrier for long distance transport. A possible source for siliciclastics could have been the Hazeva and Hordos sands that were eroded from elevated areas near the coast (Fig. 8.2c). There is no indication for clastic accumulation at the base Messinain level inland. However, evidence from several wells recently drilled in the Afiq Canyon offshore indicates some coarse clastic deposition.

8.2.5 The top-Messinian-evaporite canyon system and the Pliocene gas-bearing sands

A latest Miocene to Early Pliocene lowstand, associated with erosion of the Messinian evaporites is identified in the proximal part of the Afiq Canyon inland (Druckman et al., 1995b, 5th erosion in figure 4). Druckman et al. (1995b) noted that this lowstand is followed by fluvial erosion and flooding of the entire Mediterranean with fresh and brackish water during the so-called 'Lago Mare' event (Rouchy and Saint Martin, 1992). In the Levant Basin the top Messinian erosion is depicted by the Violet seismic horizon (Fig. 8.2d).

A short-term episode of marine deposition of the lowermost Pliocene was followed by another significant lowstand at 4.37 m.y. (Fig. 4.4). Deepwater sands that are found above the Mavkiim Formation in the offshore distal part of the Afiq canyon are separated from the
evaporites by several tens of meters thick layer of hemipelagic marl; and are therefore related to the second, Early Pliocene lowstand of 4.37 m.y.

The thick accumulations of deepwater turbiditic sands, termed the Lower Member of the Yafo Formation or the "Noa Sand" form the reservoir of the biogenic gas in the Noa, Mari and Gaza Marine fields (Fig. 1.2). These prolific gas sands represent a basin floor fan environment within the Afiq and El Arish Canyons. Some of the sands form large mounds up to 250 m in thickness. The formation of the mounds is related to postdepositional, remobilization of the deepwater sand, due to fluid injection in overpressure conditions (Oats, 2001; Frey-Martinez et al., 2007).

A likely source for the Lower Pliocene siliciclastics is the Miocene Hazeva sandstone that was exposed at that time and was eroded and transported to a short distance from the Negev and northern Sinai area (Fig. 8.2d). Another possible source is 'Nilotic' sands that were transported from the Nile Delta by longshore currents and eventually trapped and transported offshore in the Afiq canyon (Fig. 8.2d). The Yafo sands were so far encountered only in the southern part of the basin. This may be explained by the lack of adequate source of siliciclastics in the northern area during Pliocene times.

9. Hydrocarbon Potential

9.1 Trap Types and Hydrocarbon Plays

The integrated analysis of the Levant Basin presented in this report indicates a high potential for hydrocarbon accumulation. Each of the main stratigraphic intervals described and mapped in this report contains various types of potential structural and stratigraphic traps. Suggested hydrocarbon plays within each interval are shown in Figure 9.1 and are briefly described at the following sub-sections.

9.1.1 Permian to Middle Jurassic interval

This period is characterized by shallow-marine deposition and the formation of extensive graben and horst systems across the Levant region. The main hydrocarbon plays suggested for this interval are Triassic and Jurassic fault-related traps (Fig. 9.1). Alluvial fans (Erez Conglomerate type) and other coarse-grained elastic deposits may be found near the main faults or within the Triassic and Jurassic grabens. Buried hill-type traps or carbonate buildups may be found at the tops of the Triassic- Lower Jurassic highs, sealed by fine-grained Upper Jurassic and
Fig 9.1- Schematic section through the Levant Basin showing potential hydrocarbon traps in the Phanerozoic fill. Pliocene gas discoveries in the southern part of the basin and high-grade oil shows in the Mesozoic section indicate the existence of biogenic and thermogenic petroleum systems.
Cretaceous strata. These features, which are deeply buried within the basin, are located at a shallower depth (5-6 km) near the eastern margin.

9.1.2 Middle Jurassic to Turonian interval

This period is characterized by shallow-marine conditions during its early part, followed by the deposition of carbonate platforms on the margin and deepwater turbidites in the basin during the passive margin stage. The main hydrocarbon plays suggested for this interval are Lower Cretaceous deepwater fans (Fig. 9.1). The Early Cretaceous Gevara'am canyon was part of a submarine canyon and channel system incised on the continental slope in the southeastern Levant margin. Several sand layers that where deposited as deepwater fans at the distal part of this system were penetrated by offshore wells (e.g. Yam-2 and Yam West-1) (Gardosh, 2002). Similar sand bodies can be expected in other parts of the basin. It should be noted that Early Cretaceous sand bodies pre-dated the Syrian Arc deformation stage and therefore, are equally distributed on highs and lows.

The early Middle Jurassic period is characterized by the accumulation of oolitic shoals and fine-grained carbonate debris. These may be considered as an additional play type for this interval. Light oil shows were discovered in the Yam-2 and Yam Yafo-1 wells in shallow-marine, Middle Jurassic reservoirs trapped within Syrian Arc folds. Likewise the Lower Cretaceous fans, the distribution of Middle Jurassic porous intervals is not controlled by Syrian Arc deformation.

9.1.3 Oligocene to Lower Miocene interval

This period is characterized by contractional deformation and uplift that was followed by intense erosion and transport of clastic sediments into the basin. An extensive canyon and channel system developed during the Oligocene throughout the eastern part of the Levant Basin (see chapter 8.2.1). The main hydrocarbon play suggested for this interval consists of Oligocene channel-fill and deepwater fans (Fig. 9.1). Potential taps are channel-fill units interpreted on the seismic data in up-dip position (confined setting), and basin floor fans interpreted on the seismic data at the distal end of the canyons (non-confined setting) (Fig. 8.3). Some of the fans may be found in structurally favored locations within Syrian Arc II type folds.

A unique feature that developed during this period is found at the southern part of the Jonah Ridge (Figs. 5.11, 9.2). The seismic data show a large, layered mound located above a chaotic package interpreted as a Cretaceous to Early Tertiary volcanic cone, which is superimposed on a Triassic-Jurassic high (Fig. 5.11). This mound is interpreted as an atoll of possibly upper Early Miocene age, equivalent to the Ziqlag reef and shelf carbonates found on
the margin. This and other Miocene carbonate buildups that may be found in the basin, are suggested as an additional play in the Oligocene to Lower Miocene interval (Fig. 9.1). A similar carbonate buildup was likewise interpreted by Aal et al. (2000) in the ultra-deepwater of the Nile Delta, offshore Egypt.

9.1.4 Middle to Upper Miocene interval

This period is characterized by the continuation of contractional deformation, erosion and sediment transport into the basin (see chapter 8.2.1). The main hydrocarbon play suggested for this interval consists of Miocene channel-fills and deepwater fans similar to the above described Oligocene play (Fig. 9.1). Combinated, stratigraphic and structural traps are expected to be found. Deepwater channels at the base of the Messinian salt are an additional play that should be considered in this interval (Fig. 9.1)

9.1.5 Pliocene interval.

Sediment transport into the basin through a submarine canyon and channel system took place during the Early Pliocene (see chapter 8.2.1). Pliocene deepwater fans charged with biogenic gas were discovered in the Afiq Canyon. This play should be further explored in other Pliocene canyons, particularly in the southern part of the basin (Figs. 8.2d, 9.1).

9.2 Source Rocks and Petroleum Systems

An important aspect of the Levant Basin hydrocarbon potential is the distribution and maturation level of source rocks. Producing fields and hydrocarbon shows found in the basin and on its margins indicate the existence of two types of petroleum systems: biogenic and thermogenic. The organic rich shales of the Plio-Pleistocene, Oligo-Miocene and Eocene all have biogenic gas potential (McQuilken, 2001). The Middle Miocene Qantara Formation is considered by Dolson et al. (2002) as a major Tertiary source in the offshore Nile Delta. The Sadot and Shiqma gas fields (Fig. 1.2) found in the southern coastal plain are probably associated with an Oligo-Miocene source of biogenic gas. The origin of the gas found in the Noa, Mari and Gaza Marine fields is considered to be Miocene and Pliocene organic rich shale, although a particular source rock for the Pliocene biogenic gas could not be explicitly established (Feinstein et al., 2002). The great thickness and wide distribution of the hemipelagic Tertiary section, however, suggest good potential for biogenic gas generation throughout the Levant Basin.

Mesozoic source rocks comprise thermogenic petroleum systems. The Upper Cretaceous, organic rich marl of the Mount Scopus Group have excellent source properties in the inland part
of Israel and generate oil and thermogenic gas in the Dead Sea basin (Tannenbaum and Aizenshtat, 1985). Thermal maturity modeling shows that the Upper Cretaceous section reaches maturation within the Levant Basin at depths greater than 4 km (Gardosh, 2002). Therefore, Senonian strata should be considered as a potential source for oil and thermogenic gas in the deep part of the basin.

The Lower Cretaceous Gevar'am shale has source rock properties and may have generated oil and thermogenic gas where maturity is reached (McQuilken, 2001). The Middle Jurassic Barnea Formation is the source of oil in the Helez field, onshore (Fig 1.2) (Bein and Soffer, 1987). The organic rich limestone of the Barnea Formation was so far penetrated in the southern coastal plain. If the organic facies of the Barnea Formation extends into the basin (Fig. 4.5) it may serve as an important source of hydrocarbons.

Lower Jurassic and Triassic continental to shallow-marine rocks were deposited throughout the Levant area during the Neotethyan rifting stage. Bein et al.(1984) found source properties in this section in various deep, onshore wells. It is assumed that the organic content in Lower Jurassic and Triassic strata was higher in the paleograbens offshore, where lacustrine to deeper marine conditions may have prevailed. High-grade oil shows found in the Middle Jurassic limestone in the Yam-2 and Yam Yafo-1 well are probably related to these source rocks.

The timing of oil generation of the Early Mesozoic rock units is not well established. McQuilken (2001) estimated that Lower to Middle Jurassic source rocks are presently in the peak oil window to just within the gas window offshore. However, Gardosh (2002) estimated that the Triassic rocks reached the maturity window in Late Jurassic to Middle Cretaceous time. The later results further suggest that the primary migration of Early Mesozoic hydrocarbons may have taken place prior to the onset of Syrian Arc folding phase, and therefore they should be found in Early Mesozoic fault blocks and stratigraphic traps.

Recently discovered oil in the onshore Meged wells, located northeast of Helez (Fig. 1.2), was related to Silurian source rocks (www.givot.co.il). This is the first occurrence of such oil in the Levant region. It may be hypothesized that Silurian source rocks have been preserved, and generated hydrocarbons within Neotethyan rift structures such as the Judea Graben onshore and the deep-seated grabens offshore (Fig 7.1).

In summary, a wide spectrum of biogenic and thermogenic petroleum systems, ranging in age from Paleozoic to Plio-Pleistocene is found in the Levant Basin. The situation offshore Israel is probably similar to that offshore the Nile Delta where deep structures serve as focal points for vertical hydrocarbon migration, resulting in a mix of biogenic and thermogenic gases in shallow structural levels (Dolson et al., 2002, Feinstein et al., 2002).
10. Summary

Modern, geophysical data yielded new information on the structure and stratigraphy of the Levant Basin, offshore. The integration of these data with the vast amount of information from the Levant margin and inland part of Israel enables the reconstruction of a regional geologic scheme. Three distinct tectonic stages in the evolution of the region are documented:

(a) Rifting stage
(b) Post-rift passive margin stage
(c) Convergence stage

Rifting in the Levant region is related to the breakup of the Gondwana plate and the formation of the Neotethys Ocean system. Four extensional pulses are postulated: in the latest Paleozoic; Middle Triassic; Late Triassic; and Early Jurassic. These pulses resulted in normal faulting in the range of several kilometers, magmatic activity and the formation of NE-SW oriented graben and horst systems. Four basement structures associated with the Early Mesozoic extension are found in the deep parts of Levant Basin (from east to west): the Yam, Jonah, Leviathan, and the Eratosthenes highs. The main structures inland are the Gevim, Gaash and Maanit highs and the Hilal, Judea and Asher grabens. The rifting in the Levant area reached an early magmatic stage. Although magmatic intrusion and stretching of the crust took place, no indications for sea-floor spreading and emplacement of new oceanic crust are found. Continental to shallow-marine depositional environments were dominant in the Levant region during the rifting stage.

The post-rift stage is associated with cooling and subsidence that was probably more intense within the basin than on the eastern margin. The Late Jurassic to Middle Cretaceous section records the gradual formation of a passive-margin profile and the subsequent development of a deep marine basin bordered by a shallow-marine shelf. The passive-margin stage is characterized by recurring cycles of marine transgression and regression associated with relative sea-level changes. These are reflected by marine onlaps, numerous unconformity surfaces, accumulation of carbonate platforms on the margin and mass transported deposition in the basin.

The convergence stage is related to the closure of the Neotethyan Ocean system and the motion of the Afro-Arabian plate towards Eurasia. In the Levant region this motion is manifested by large-scale contractional deformation of the Syrian Arc fold belt. Two contractional phases are observed. A Late Cretaceous Syrian-Arc I phase is characterized by inversion of older normal faults and the development of asymmetric, high-amplitude folds that are found mostly near the eastern margin and further inland. An Oligo-Miocene Syrian-Arc II
phase is characterized by the formation of low-amplitude folds throughout the basin, and uplifting and tilting of marginal blocks, at the eastern part of the basin and further inland.

Tertiary tectonic activity and eustatic sea-level falls caused intense erosion and accumulation of several kilometers thick basin-fill. Fluvial systems flowed on the exposed shelf from elevated areas in the east. Submarine turbidity currents incised the upper slope and transported siliciclastics further into the basin. The Oligo-Miocene and Pliocene drainage system includes, from south to north, the Afiq, Ashdod, Hadera, Casarea, Atlit and Qishon canyons. The Afiq and Ashdod canyons are highly erosive and are incised several hundred meters in the continental slope. Other canyons are less erosive and their flow pattern was for the most part controlled by the topography of the paleoslope. The Tertiary drainage system was shut off temporarily during the Messinian salinity event and turned on again during post-evaporites, latest Miocene to Early Pliocene lowstands.

The variety in tectonic styles and depositional patterns may provide favorable trapping conditions for hydrocarbons in the Levant Basin. Potential structural traps are associated with the extensional rift structures and the contractional Syrian Arc folds. Stratigraphic traps are associated with Triassic-Middle Jurassic shallow-marine, carbonate and siliciclastic reservoirs and Cretaceous and Tertiary deepwater turbidite systems.

Shallow gas discoveries in Pliocene sands and high-grade oil shows found in the Mesozoic section indicate the presence of source rocks and appropriate conditions for hydrocarbon generation in both biogenic and thermogenic petroleum systems. The size, depth and trapping potential of the Levant Basin supports the conclusions that large quantities of hydrocarbons can be found offshore Israel.

This study was focused on the stratigraphy, structure and tectonic evolution of the Levant Basin. It is recommend to conduct specific studies on the following topics: (a.) further analysis of the Tertiary drainage systems in wells, outcrops and 3D seismic data; (b) further analysis of the Middle Jurassic to Middle Cretaceous reservoir rocks and depositional environments, (c.) 2D and 3D structural reconstruction of the basin; (d.) modeling the thermal history and migration paths for hydrocarbons; and (e.) study of Direct Hydrocarbon Indicators (DHI) in 2D and 3D seismic data.
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