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Tectonic and Eustatic Controls on the Evolution of
the Maldivian Carbonate Platform

by

Andrei Victorovich Belopolsky

A Thesis Submitted in Partial Fulfillment of the Requirements for
the Degree
Doctor of Philosophy

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ABSTRACT

TECTORIC AND EUSTATIC CONTROLS ON THE EVOLUTION OF THE MALDIVE CARBONATE PLATFORM

by

Andrei Belopolsky

The Maldive Archipelago in the equatorial Indian Ocean is only the uppermost part of a more than 3-km thick carbonate platform. The Maldive platform contains a 50 Ma-long sedimentation record and has a relatively simple tectonic history. The interpretation of 6000 km of Shell 2-D seismic data and information from two industry and three ODP wells was the basis for the reconstruction of the platform evolution and assessment of controls on platform development.

The evolution of the Maldives platform was essentially twofold. During the first stage (Eocene-early Oligocene), tectonic control played the dominant role in the establishment and geographic distribution of shallow water carbonates. A series of shallow water carbonate platforms were formed in the early Eocene on basement highs separated by two deep, narrow, and continuous graben systems. The platforms aggraded and backstepped in the Eocene and early Oligocene in response to relative sea level rise driven mostly by tectonic subsidence.
The second stage of the platform evolution (late Oligocene-Quaternary) was predominantly controlled by sea level fluctuations. A significant fall in sea level at the early-late Oligocene transition, with a magnitude possibly up to a 100 m, was recorded in the paleo-bathymetry of the Shell ARI-1 well. In the late Oligocene and early Miocene, the platforms first aggraded, partially drowned, and later backstepped in response to a substantial long-term sea-level rise. At the end of the early Miocene, a series of aggrading flat top carbonate banks, a small remnant of the Eocene-Oligocene neritic carbonate system, were established on the periphery of the central basin, the predecessor of the modern Inner Sea of the Maldives. During the middle Miocene, the bank margins prograded for 10-15 km. The progradation was driven by five complete sea level cycles, with each cycle represented by a relative sea-level fall and a subsequent rise.

The reconstructed late Oligocene - middle Miocene relative sea level history of the Maldives corresponds well with the newly-published ice-volume record based on the temperature-corrected benthic foraminifera oxygen isotope data. The late Oligocene-middle Miocene depositional geometries of the Maldives platform appear to have recorded eustatic sea-level fluctuations.
ACKNOWLEDGMENTS

The four years I spent at Rice were truly enjoyable. This is because of the excellent academic and social environment at both the Department of Geology and Geophysics and at Rice University as a whole. The second part of my term as a graduate student at Rice was mainly spent at the Shell Bellaire Technology Center in Houston. It became my "second department", and I greatly enjoyed learning and interacting with people there.

First and foremost, my thanks go to my research advisor, André Droxler, who became my teacher and my friend. André never stops sharing his ideas and his passion for science, and his interaction with students extends far beyond the lab environment. In my opinion, André is the kind of advisor any graduate student could wish for.

André is responsible for introducing me to the Maldives. I feel privileged to have worked on the Shell Maldives data set that has unsurpassed quality and volume. Years after I started this project, I am just as excited about this data as I was when I saw the first seismic line from the Maldives. I am grateful to the Royal Dutch/Shell Company, in particular Peter Suessli, who provided our research group with the seismic and well data from their exploration campaign. I also express my sincere gratitude to the Maldives government that authorized the data release. The funding for this project was provided by the National Science Foundation Ocean Sciences Grant, the Mills Bennet Fellowship from Rice University, and the SEPM Grant-in-Aid.
I would like to thank the geologists who served on my thesis committee: John Karlo, Mitch Harris, Richard Gordon, and Peter Vail. Each of them took an informal approach to supervising my research and I greatly enjoyed interacting and learning from them. I also thank Joan Strassmann from the Department of Ecology and Evolutionary Biology who bravely agreed to serve on the thesis committee.

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This project would not be possible without the support of my family. I thank my parents, Victor and Natalia, my brother Artem and my sister Alisa, for their unconditional love and support. To them I dedicate this thesis.
Reflecting how powerful an agent with respect to denudation, and consequently to the nature and thickness of the deposits in accumulation, the sea must ever be, when acting for prolonged periods on the land, during either its slow emergence or subsidence; reflecting, also, on the final effects of these movements in the interchange of land and ocean-water, on the climate of the earth, and on the distribution of organic beings. I may be permitted to hope, that the conclusions derived from the study of coral formations, originally attempted merely to explain their peculiar forms, may be thought worthy of the attention of geologists.

Charles Darwin, The Structure and Distribution of Coral Reefs, 1842
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PREFACE

My dissertation includes the results of a study on the evolution of the Maldive carbonate platform in the equatorial Indian Ocean. The dissertation consists of two parts, preceded by a general introduction and followed by a general conclusion. Part I entitled "Tectonic and eustatic controls on the evolution of the Maldive carbonate platform: a 50-Ma-long case history", is dedicated to the overall evolution of the Maldive platform from the moment of its inception until the present time. Because of the significant volume and numerous oversized seismic lines and maps that constitute Part I, it would not be feasible to publish it as a research paper in an academic journal. The intention is to publish Part I of the dissertation in a form of an atlas that will include large plates displaying maps and both uninterpreted and interpreted seismic sections. In addition to the figures included in Part I, the atlas will comprise a number of other characteristic seismic sections interpreted in this study but not shown in this dissertation. Also, an abridged version of Part I is planned to be submitted to the AAPG Bulletin in a form of a research article.

Part II of the dissertation is entitled "Carbonate bank response to the middle Miocene sea-level fluctuations in the Maldives (equatorial Indian Ocean)". This part is focused on the extraction of the relative sea level record from the middle Miocene prograding sequences. Part II presents a stand-alone research paper and will be submitted to the Journal of Sedimentary Research, Stratigraphy and Global Studies section. A small
overlap between the Parts I and II was unavoidable and some of the figures were used interchangeably.

During the course of my Ph.D. study, I was also involved in two other research programs unrelated to the Maldives the study, published as two first-author papers. The first publication, entitled “Uppermost Pleistocene transgressive coralgal reefs on the edge of the South Texas shelf: analogs for reefal reservoirs buried in siliciclastic shelves”, was co-written with André Droxler and presented the results of the seismic stratigraphic study of Southern Bank, a coralgal reef partially exposed on the seafloor on the edge of the South Texas shelf offshore Corpus Christi, Texas. This paper was published in 1999 by the Gulf Coast Section SEPM Foundation in the proceedings of the Nineteenth Annual Research Conference “Advanced reservoir characterization for the twenty-first century” (GCSSEPM, Houston, TX, p. 41-50). The other research project was focused on estimating the Caspian Sea region petroleum potential. The resulting paper was published in 1998 as a Special Publication “Geology and petroleum potential of the Caspian Sea region” by the James Baker III Institute for Public Policy (Rice University, Houston, TX, 33 p.). Manik Talwani and Diane Berry were co-authors of this paper. This publication was a part of the multidisciplinary research study “Unlocking the Assets: A Political, Economic, and Cultural Analysis” conducted by the Baker Institute at Rice University. Members of my thesis committee have advised me not to include these two research papers in the dissertation volume to conserve the wholesomeness of the Maldives research.
GENERAL INTRODUCTION

Shallow-water carbonate sediments deposited in tropical and subtropical settings form thick and spatially extensive accumulations referred to as “carbonate platforms”. Carbonate platforms typically have a life span of millions to tens of millions of years, and their birth, growth, and demise are governed by a combination of factors such as tectonics, eustasy, environmental conditions, and climate. Variation of these different factors through time are archived in the platform sedimentary record and, once extracted from it, enhance our knowledge of the Earth history and our understanding of sedimentary processes.

Deposition of sediments is largely controlled by the interplay between the accommodation space and sediment supply. New accommodation space, or space available for sediment accumulation, is created or destroyed by tectonic subsidence or uplift, and eustatic sea-level fluctuations. The sedimentary system responds to changes in the accommodation space and sediment supply, and these changes are recorded in the stratal geometries of the system.

The recognition and interpretation of characteristic patterns of stratal geometries is one of the fundamentals of sequence stratigraphy. Historically, sequence stratigraphic studies were focused on the siliciclastic systems. The original ‘sea slug’ model of Vail et al. (1977) withstood the test of time and has been used as a foundation for numerous
case-specific modifications. The application of this model to the carbonate systems, however, is arguable because of the major differences in the processes of sediment formation and accumulation between the two systems. The simple re-labeling of lithologies and facies in the ‘sea slug’ model (e.g. Sarg, 1988) does not take into account some of the major differences that exist between siliciclastic and carbonate systems.

Carbonate sediment production occurs by biogenic and biochemical processes locally on the bank top, often referred to as the ‘carbonate factory’. The in situ sediment production makes the carbonate system dramatically different from the siliciclastic system, in which almost all deposited sediments are transported to the basin by rivers. Because carbonate sediment production on the bank top is confined to the euphotic zone, the rate of production is highly sensitive to the relative sea level fluctuations. The environmental control (climate, circulation patterns, and nutrient supply) also has a strong influence on the carbonate sediment production. If carbonate production is impaired or shut-off for an extended period of time, a carbonate platform can be terminated and drown – a phenomenon which does not exist in the silicilastic settings.

Carbonate platforms possess characteristic stratigraphic geometries that are recognized in outcrops and on seismic profiles (Fig. 0.1). The development of these geometries is controlled by the interplay of two factors: the rate of carbonate production and growth (G), and the rate of new accommodation space creation (A). The measured growth potential of modern shallow-water carbonates exceeds any known rate of a sea-level rise. This observation led to the postulation of the “paradox of drowned reefs and carbonate
Figure 0.1. Common geometries of carbonate platforms (modified from Schlager, 1992). A=rate of new accommodation space creation, Gp=growth potential of the platform interior, Gr=growth potential of the platform rim.
platforms" (Schlager, 1981). The scaling of sedimentation rates in geological time, however, explains the long-term process of drowning (Schlager, 1999).

Within the carbonate platform itself, the rate of carbonate production may vary geographically. Platform rims typically display higher growth rates than platform interiors. The rims maintain their tops close to sea level while the platform interior is commonly submerged within the euphotic zone in a few to tens of meters of water.

When the rate of carbonate production exceeds the rate of accommodation space creation, the platform aggrades (vertical growth) and progrades (lateral growth) at the same time (Fig. 0.1A). This situation corresponds to the highstand systems tract in the sequence stratigraphic model. When the platform top is flooded, and carbonate factory operates at full strength and the excess of produced material is exported offshore. In the transgressive systems tract, when the relative sea level is rising, the platforms may form different stratigraphic geometries depending on the A/G ratio. When the rate of the new accommodation space creation exceeds the growth potential of the platform, the platform backsteps trying to "keep up" with the rising relative sea level (Fig. 0.1B). This geometry is recognized best in rimmed carbonate platforms where the up-slope migration of elevated margin buildups easily catches the eye.

If the rate of the new accommodation space creation equals the growth potential of the platform, the platform aggrades vertically maintaining its flat top at or near sea level (Fig. 0.1C). Another scenario includes the situation when the rate of the new accommodation space creation is equal or smaller than the growth potential of the platform rim but exceeds the growth potential of the platform interior (Fig. 0.1D). This
leads to the aggradation of the platform rim and the drowning of the platform interior. The result is the characteristic “empty bucket” geometry (Kendall and Schlager, 1981; Schlager, 1981) with a stiff rim and an interior partially filled with soft carbonate mud.

In the situation when the rate of the new accommodation space created greatly exceeds the growth potential of both platform rim and platform interior, the carbonate platform would drown (Fig. 0.1E). Shallow-carbonate production ceases and pelagic sediments are deposited above the platform top. Drowning unconformities are common in the geological record and may be misinterpreted as created by exposure above sea level. Exposure of the platform top, however, is not required for platform drowning (e.g. Schlager, 1998, 1999). Drowning unconformities are pronounced unconformities recognized both in the outcrop and on seismic sections. Submarine erosion is common on the drowned surfaces. The platform interior submerged below the euphotic zone may be subjected to strong currents that sweep and erode the platform top (Schlager, 1998). This makes it more difficult to distinguish a drowning unconformity from subaerial exposure surfaces, in particular on seismic profiles alone.

In the event of a relative sea-level fall, or a reduction of the accommodation space, the platform top becomes subaerially exposed (Fig. 0.1F). The carbonate factory either shuts down completely or migrates downslope. The exposed platform top is subject to meteoric diagenesis. The carbonate sediments become subject to rapid cementation, erosion, and dissolution. Prolonged exposure lead to the development of karst topography. Basinward migration of the platform, or downstepping (Fig. 0.1F) may also occur. The new area of shallow-water carbonate production and accumulation largely
depends on the depositional profile. In carbonate platforms with a ramp or gentle slope morphology, the area of the new downstepped platform may not be reduced significantly while on the steep-slope platform the new shelf will be narrow (Kendall and Schlager, 1981; Hanford and Loucks, 1993). This process results in the characteristic "forced-regressive" geometry of a downstepping wedge attached to the platform slope (Posamentier et al., 1992; Handford and Loucks, 1993; Hunt et al., 1993). In the event of repetitive sea-level falls punctuated by stillstands, a series of downstepping wedges will form.

This study is focused on the Maldives isolated carbonate platform in the equatorial Indian Ocean. The Maldives platform is characterized by very large size (800 by 130 km) and is the second largest (after the Bahamas) modern isolated carbonate platform. The Maldives platform comprises a continuous since the early Eocene record of shallow carbonate sedimentation. The thickness of the sedimentary section exceeds 3 km.

It is surprising that the Maldives platform received so little attention compared with the Bahamas and other, smaller carbonate platforms, and considering the vast amount of data available in the Maldives. Only two research groups of geologists have worked previously in the Maldives. André Droxler at Rice University has been involved in the research in the Maldives for well over a decade. Under his supervision, Olivier Aubert completed his Ph.D. research in 1994. Their work included the interpretation of the Elf seismic data set calibrated by the well data from ODP, Elf and Shell and was published as two research papers (Aubert and Droxler, 1992, 1996).
Ed Purdy has also been involved in research in the Maldives for the past two decades. In 1993, together with George Bertram, he published the study "Carbonate concepts from the Maldives, Indian Ocean". The work of Purdy and Bertram was based on the interpretation of Elf seismic lines and Elf NMA-1 well. As a result of the ODP Leg 115 research, other programs in the Maldives have focused on the Plio-Pleistocene periplatform sediments (Droxler et al., 1990; Malone et al., 1990) and the initial establishment of the carbonate system in the Eocene (Nicora and Premoli Silva, 1990).

The release of the entire seismic and well data sets from the Shell exploration campaign opened a new opportunity to conduct a detailed and three-dimensional study of the platform evolution. The excellent quality and the unprecedented volume (6000 km), for an academic research project, of the seismic data set facilitated the creation of the detailed regional model presented in this study.

The seismic expressions in the Maldives of the carbonate stratal geometries ground-truthed by several wells are undoubtedly the best examples available to the research community at the present time. All geometries typical for carbonate platform evolution (Fig. 0.1) are found on the Maldive seismic sections.
PART I.

TECTONIC AND EUSTATIC CONTROLS ON THE EVOLUTION OF THE
Maldive Carbonate Platform: A 50 Ma-long Case History

ABSTRACT

The Maldive Archipelago in the equatorial Indian Ocean is only the tip of a more than 3-km thick carbonate platform which was established in the early Eocene and has thrived ever since. One of the largest isolated carbonate platforms of the Earth, the Maldives contain a 50 Ma-long sedimentation record and have a relatively simple tectonic history. In this study, reconstruction of the entire platform evolution and assessment of various controls on platform development are based on the interpretation of 6000 km of Shell 2-D seismic data that covers the central part of the platform, and information from two industry and three ODP wells.

The evolution of the Maldives platform was essentially twofold. During the first stage (Eocene-early Oligocene), the basement structure and its tectonic control played the dominant role in the establishment and geographic distribution of shallow-water carbonates. A series of shallow-water carbonate banks were established in the late early Eocene on basement highs separated by two deep, narrow, and continuous graben systems. The grabens displacing the volcanic basement of late Paleocene age served as
deep seaways. The platforms aggraded and backstepped in the Eocene and early Oligocene in response to the relative sea level rise driven mostly by tectonic subsidence. Movement along the graben faults diminished through time and ceased by the end of the early Oligocene.

The second stage of the platform evolution (late Oligocene-present) was predominantly controlled by eustatic sea level fluctuations. The quiescent tectonic regime during this time is explained by the Maldives position on the rigid Indian plate, away from plate boundaries and stress fields. A significant fall in sea level, possibly of magnitude up to 100 m, at the early-late Oligocene transition can be inferred from the paleo-bathymetric information of the Shell ARI-1 well. In the late Oligocene and early Miocene, the banks first aggraded, partially drowned, and later backstepped in response to a substantial long-term rise in sea level. At the end of the early Miocene, a series of flat top carbonate banks, a small remnant of the Eocene-Oligocene neritic carbonate system, were established on the periphery of the central basin, the predecessor of the modern Inner Sea of the Maldives. During the middle Miocene, the bank margins prograded for 10-15 km on both sides of the central basin in response to five complete sea-level cycles. The reconstructed late Oligocene-middle Miocene relative sea level history of the Maldives corresponds well with the newly-published ice-volume record based on the temperature-corrected benthic foraminifera oxygen isotope data.

The progradation of some of the bank margins continued until the early Pliocene while other banks aggraded or drowned. The late Pliocene was characterized by a
basinward shift in facies related to a sea-level fall caused by the onset of the main Northern Hemisphere glaciation. The Pliocene exposed banks were periodically flooded and exposed by Quaternary high frequency sea level fluctuations. The reestablishment of the carbonate production on the antecedent karst topography caused the change from the early Pliocene flat-top bank morphology to the modern atoll configuration.
INTRODUCTION

Reconstruction of the entire evolution of any large sedimentary system presents a formidable yet highly rewarding task. In addition to providing an understanding of the distribution of the sedimentary lithologies, facies and mineral resources within the basin, the system's history also contains information on changes in regional and global plate tectonic, environmental, eustatic and climatic conditions. Unfortunately, few seismic data sets with comprehensive data coverage are available for large regional studies. Insufficient data may result from gaps in spatial coverage and incomplete subsurface information when only part of the section is ground-truthed by drilling.

In the presented study, a dense seismic grid and information from wells provide a unique opportunity to reconstruct the evolution of the Maldives, a large carbonate platform, from its inception to its present state. By being able to image the basement structure and fault patterns, the degree of tectonic control on the long-term evolution of the platform can be assessed. The interpretation of stratal geometries, supported by the lithologic, biostratigraphic, and paleowater depth information from wells, sheds light on the platform development in response to the relative sea level fluctuations. The goals of this study were: 1) to image the basement structure and show the tectonic control on the establishment of the carbonate platforms, 2) to reconstruct Maldives platform morphology at different stages of its development, 3) to determine the role of tectonic and eustatic
factors in the evolution of the Maldive platform and, 4) to compare the Maldives Oligocene-middle Miocene relative sea level record with global sea level proxies such as deep-water benthic foraminifera oxygen isotope records and coastal onlap curves.
BACKGROUND

General Setting of the Maldives

The atolls of the Maldives archipelago form the central and largest part of the Laccadive-Chagos atoll chain in the equatorial Indian Ocean (Fig. I.1). The north-south trending Laccadive-Chagos chain extends from the southwest coast of India to south of the equator and is composed of low-lying coral atolls. The main frame builders of the Maldivian reefs today are scleractinian corals, and green algae _Halimeda_ and _Tydeania_ are the most important producers of carbonate sand and mud (Ciarapica and Passeri, 1993). The Maldives archipelago consists of 22 large atolls whose size ranges from a few km to tens of km in diameter (Fig. I.2). The atolls are arranged in clusters separated by deep channels. Ihavandifillus, the northernmost Maldivian atoll, is separated from the Laccadive islands (India) by the deep Eight Degree Channel (Fig. I.1). The southernmost Maldivian atoll, Addu, is located south of the equator and is separated from Suvadiva, one of the largest atolls in the world, by the Equatorial Channel. Suvadiva, in its turn, is divided from the central group of atolls by the One and Half Degree Channel. The shapes of the atolls vary from circular to elongate in map view, and numerous smaller atolls called "faros" are commonly present within the lagoons of the large atolls, and in places form the rims of the large atolls. The depth of the lagoons ranges from 31 to 82 m (Purdy
Figure I.1. Atolls of the Maldivian Archipelago, central equatorial Indian Ocean. Black areas represent atoll islands, grey areas indicate atoll lagoons.
Figure 1.2. Central Maldives atolls and bathymetry map. Locations of ODP Sites 714, 715 and 716 and industry wells ARI-1 and NMA-1 are also shown.
and Bertram, 1993) and tends to increase from north to south. Although the archipelago extends for 867 km from north to south, the island area is only 298 km². Approximately 1,200 individual islands exist, but only 200 of them are populated.

In the central part of archipelago, the large atolls are arranged in two parallel north-south-trending chains separated by the Inner Sea (Fig. I.2). Several large drowned flat-top banks complete the “broken” segments of the double chain of atolls. Fuad Bank between Horseburgh and Ari atolls, with its top submerged in 250 m of water, is an example (Fig. I.2). The water depth of the Inner Sea ranges between 200 and 500 m. The combined width of the platform (atolls and the Inner Sea) locally is up to 130 km. The large size and the central basin make the Maldives dramatically different from many other oceanic atolls and guyots. Isolated carbonate atolls and guyots typically possess a simple aggrading geometry and steep erosional margins (e.g. Sarg, 1988). As demonstrated below, the large dimensions of the Maldive platform and the relatively shallow central basin facilitated the formation of depositional geometries similar to those observed on many passive margins.

The flanks of the atolls on the Inner Sea side are characterized by a steep gradient down to approximately 150 m, after which point they become gentle. In contrast, steep gradients characterize the slopes of oceanward flanks of the atolls. The oceanward slopes, typical for erosional bypass margins, quickly reach water depths in excess of 2000 m (Fig. I.1). Downslope mass-wasting deposits, such as slumps, are readily recognized on seismic
profiles at the base of the margins (Aubert, 1994; Aubert and Droxler, 1996; Purdy and Bertram, 1993).

The Maldives have a tropical oceanic climate with high humidity and stable temperatures throughout the year. In January, the northern part of the archipelago is subjected to westerly surface currents driven by the Indian north-easterly monsoon winds. The southern part is affected by easterly currents that are part of the Equatorial Counter Current. In July, most of the archipelago has easterly currents due to the Indian south-westerly monsoons; however, the strength of the current is weak (UNEP/IUCN, 1988).

**Tectonic Setting of the Maldives**

The area of the equatorial Indian Ocean presently is characterized by unusual deformation (Gordon et al., 1998). Royer and Gordon (1997) proposed that the Indo-Australian plate was not one homogeneous plate but consisted of three distinct rigid plates separated by multiple zones of deformation. These zones of deformation represent diffuse plate boundaries that separate the Indian, Australian, and a newly defined Capricorn plate (Fig. I.3). The present day position of the Maldives is within the rigid Indian plate, away from the diffuse plate boundaries. This explains the relative absence of seismic activity in the vicinity of the Maldives at present (Fig. I.4). In contrast, Chagos Bank, located at the southern end of Laccadive-Chagos Ridge and 800 km away from the southern end of the Maldives archipelago, is characterized by high seismic activity
Figure I.3. Tectonic plate geometry and wide deformation zones in the central Indian Ocean (modified from Royer and Gordon, 1997). Plates shown include the Indian Plate (INDIA), the Capricorn Plate (CAP), the Australian Plate (AUS), the Antarctica Plate (ANT), the Somalian Plate (SOM), the Arabian Plate (ARA), and the Eurasian Plate (EURASIA). Open arrows indicate diffuse plate boundaries accommodating horizontal divergence, and filled arrows show the diffuse plate boundaries accommodating horizontal convergence. According to this model, the Maldives are positioned within the rigid Indian Plate several hundred km away from the diffuse plate boundaries. Chagos Bank is located within a deformation zone.
(Fig. I.4). This phenomenon is explained by the position of Chagos within the extensional diffuse plate boundary zone (Fig. I.3). In the western part of the stretching zone between the Indian and Capricorn plates, the deformation is mainly expressed by the right-lateral strike-slip faulting along NE-striking fracture zones (Royer and Gordon, 1997). Plate reconstructions of Gordon et al. (1998) show that the motion between the Indian and Capricorn plates started at least 20 to 18 Ma ago. The Indo-Australian plate is thought to have existed since 43 Ma (chron 20) when spreading between the Indian and Australian plates ceased in the Wharton basin (Liu et al., 1983).

**Structure and Origin of the Chagos – Laccadive Ridge**

The Chagos-Maldives-Laccadive aseismic ridge stretches for 3000 km in the north-south direction from the southwestern coast of India to south of the equator along the 73rd-degree meridian (Fig. I.4). The origin of the ridge is commonly attributed to the activity of the Réunion hot spot (Morgan, 1972, 1981; Duncan et al., 1990). The Réunion hot spot track includes the massive Deccan Trap basalts in India, the Chagos-Laccadive ridge, and the Mascarene Plateau in the southwestern Indian Ocean (Fig. I.5). The hot spot became active with the eruption of the Deccan trap basalts in India around the K-T boundary time, but the basalt eruptions probably both preceded and postdated it (Sheth and Chandrasekharam, 1997). The fixed hot spot position under the northward-moving Indian Plate later created the volcanic ridge that forms the foundation of the Laccadives,
Figure I.5. Structural and tectonic elements of the western Indian Ocean. The map shows the Réunion hot spot trail: Deccan Trap basalts, Chagos-Maldive-Laccadive volcanic ridge, Mascarene Plateau (which includes Saya de Malha, Nazareth and Cargados Carajos Banks), Mauritius, and Réunion. Open circles mark the locations of ODP Leg 115 sites, with ages of recovered basalts plotted next to them.
the Maldives, and Chagos Bank. The Mascarene Plateau which includes Saya de Malha, Nazareth, and Cargados Carajos Banks, along with the islands of Mauritius and Réunion, form the other segment of the hot spot track in the southwestern Indian Ocean. The present position of the hot spot is thought to be under the island of Réunion, where Piton de la Fournaise volcano is the surface expression of the hot spot (Freitzdorff et al., 1998). Testing the hot spot origin of the Deccan-Réunion volcanic trail was one of the main objectives of the Ocean Drilling Program (ODP) Leg 115 in 1987 (Backman et al., 1988; Duncan et al., 1990). Basement samples were recovered and dated with $^{40}$Ar/$^{39}$Ar at four sites (Fig. I.5). The age distribution was progressively younger from north to south along the trail and in agreement with the computer modeling of the plate movement (Duncan et al., 1990), thus making a strong argument for the hot spot origin of the ridge.

The Chagos-Laccadive Ridge is separated from the Mascarene Plateau by the Central Indian Ridge (Fig. I.5). The reorganization of the Central Indian Ridge occurred at 38 Ma (Patriat and Segoufin, 1988) when the ridge moved away from the hot spot at the time of plate reorganization (Burke, 1996).

Earlier hypotheses about the origin of the Chagos-Laccadives ridge included a 'leaky' transform fault (Fisher et al., 1971; McKenzie and Sclater, 1971), a microcontinent (Krishan, 1960), and a combination of the structural elements listed above (Avraham and Bunce, 1977). Milanovsky and Milanovsky (1999) attributed the eruptions of the volcanic rocks of Chagos-Laccadive Ridge and the Ninetyeast Ridge to a number
of north-south trending transform faults where the spreading had ceased and dip-slip, divergent or pull-apart movements occurred.

Burke (1996) advocates a modified scenario of the hot spot origin, in which the formation of the Chagos-Laccadive ridge is related to the activity of a different hot spot other than Réunion. He proposed that the African Plate had been at rest with the underlying mantle circulation for the past 30 Ma. Burke (1996) concluded that the Deccan Trap mantle plume source had been inactive since 30 Ma ago and is currently buried under the carbonate edifices of Nazareth or Cargados Carajos Bank. The basalts recovered in ODP Site 706 on the Nazareth Bank gave an age of 33 Ma and basalts from Texaco’s Nazareth Bank well NB-1 yielded an age of 31 Ma (Duncan and Hargraves, 1990). Rodrigues Ridge, along with Rodrigues island, Mauritius, and Réunion, form a 1000 km-long trend at an approximately right angle to the proposed Deccan hot spot track, and do not show an internal age progression. Burke (1996) thus proposed that a young intraplate hot spot, different from the one responsible for the formation of the Deccan Trap basalts and Chagos-Laccadive Ridge, created the Rodrigues-Réunion group of islands.

Sheth (1999) dismisses the mantle plume hypothesis and suggests that the creation of the Laccadive-Réunion volcanic ridge was related to a southward-propagating fracture, and that the ultimate cause of the Deccan eruptions was mega-scale lithospheric rifting. The Laccadive-Maldive-Chagos section of the track is positioned along the Vishnu fracture zone, and the Saya de Malha-Réunion segment is situated along the Mauritius
fracture zone. According to Sheth (1999), the seamount chain reflects the stress conditions in the Indian Ocean and is not related to the activity of a fixed melting point underneath a rigid continental plate.

Previous Studies of the Maldive Carbonate Platform

Early expeditions to the Maldives included bathymetric surveys as well as studies of the ecology of the coral reefs and associated fauna (see summaries in Ciarapica and Passeri, 1993, and Purdy and Bertram, 1993). A myriad of atolls in the Maldives stimulated some of the greatest scientific minds to develop the theories of atoll formation (Darwin, 1842; Agassiz, 1903). However, the low-lying atolls would provide little clue about the history of the Maldives beyond the Holocene. This situation changed when oil companies became interested in the potential for hydrocarbons in the Maldives. From 1968 to 1978, a consortium led by Elf Aquitaine held an exploration license over the Inner Sea basin and much of the archipelago. Between 1971 and 1974, Elf acquired 6750 km of 2-D deep marine and shallow water seismic data, 2700 km of which was collected in the atoll lagoons (Fig. I.6). In 1976, Elf drilled their NMA-1 (TD 2221 m) water-bearing well in the lagoon of the North Male atoll. The drilling encountered good carbonate reservoirs and a potential source rock before penetrating the late Paleocene basalts. In 1987, ODP Leg 115 drilled several sites in the equatorial Indian Ocean, including three in the Maldives (Backman et al., 1988). The NMA-1 well, the ODP sites,
Figure I.6. Elf Aquitane 1971-74 2-D seismic grid (6750 km), part of which was interpreted by Aubert and Droxler (1992, 1996), Purdy and Bertram (1993), and Aubert (1994). Locations of Elf NMA-1 and Shell ARI-1 wells and ODP Sites 714, 715, and 716 are also shown.
and a significant portion of the seismic data acquired by the Elf consortium were studied by Aubert and Droxler (1992) and Purdy and Bertram (1993).

In 1989, Shell Maldives Exploration and Production B.V. was awarded the Inner Sea Basin concession (39,953 km²). Subsequently, in 1989-90, Shell acquired and processed 6000 km of 2-D seismic data (Fig. I.7). In 1991, the commitment ARI-1 well was drilled in the central part of the Inner Sea east of Ari atoll. The well penetrated over 3300 m of porous (10-30 %) carbonates and bottomed in the weathered basalts. Immature source rocks and a lack of sealing lithologies prevented the generation and entrapment of hydrocarbons in the Maldives. In 1991, Shell Maldives B.V. was terminated and the acreage relinquished. The interpretation of the Elf seismic data tied to the Shell ARI-1 well was published by Aubert (1994) and Aubert and Droxler (1996). In 1997, Royal Dutch Shell, with the permission from the Maldivian government, released the entire seismic data set and well information to the research group at Rice University in Houston, Texas.
Figure I.7. Shell 2-D multi-channel seismic (MCS) grid and locations of Elf NMA-1 and Shell ARI-1 wells and ODP Site 716. Thick lines indicate seismic lines discussed in text.
DATA AND METHODS

Well data

The two industrial wells (ARI-1 and NMA-1) and three ODP Sites (714, 715, and 716) drilled in the Maldives provide information on the lithology, stratigraphy, and depositional environments of the sediments that accumulated in the Maldives over the last 50 Ma. This knowledge is essential for the ground-truthing of seismic interpretations. The wells were drilled in a variety of present-day depositional settings: NMA-1 in an atoll lagoon; ARI-1 and ODP 716 in the central part of the Inner Sea, and ODP Sites 714 and 715 on the oceanward slope of the modern platform. Altogether, the wells complement each other having variable penetration depth providing a comprehensive control for this study.

The NMA-1 well was drilled by Elf in 1976 in the North Male atoll lagoon in 46 m of water (Fig. I.6). The well penetrated 2106 m of Eocene to Recent carbonate sediments and 116 m of late Paleocene weathered basalts considered to be the basement (Fig. I.8, I.9). No hydrocarbon shows were encountered in the NMA-1 well that tested only water. The carbonate sediments were highly porous (15-35%), except for a tight dolomite section (porosity <5%) just above the basalts. 70 m-thick section of black organic-rich late Oligocene calcareous shales was recovered in the interval 1320-1490 m
Figure I.8. Litho- and biostratigraphy of the Elf's NMA-1 well (modified after Aubert and Droxlcr, 1996).
Figure I.9. Elf NMA-1 well drilled in the lagoon of North Male atoll and Shell seismic line E310NMA. Seismic interpretation and horizons are discussed further in the text. Uneven sea floor bathymetry causes velocity pull-up at the atolls margin. See seismic line location on Figure I.7.
(van Gils and Rubbens, 1992). The age and depositional environment of the sediments for this well were determined from cuttings and sidewall core analysis (Lehman and Pons, 1986). The record is incomplete because two significant intervals had no recovery (350-761 and 1500-1833 m). At these levels, numerous losses of circulation and caving occurred while drilling, probably related to large open and interconnected fractures and/or karstification. In addition to core and cuttings samples, the well has standard gamma ray, sonic, resistivity, spontaneous potential and density logs.

In 1991, Royal Dutch Shell drilled the ARI-1 well in the central part of the Inner Sea 14 km east of Ari atoll in the water depth of 348 m (Fig. I.7). The well was drilled to test a seismically defined dip-closed structure. ARI-1 penetrated porous (15-30% porosity) carbonates of late Eocene to Recent age and bottomed in 50 m of weathered basalts without encountering any hydrocarbon indications (Figs. I.10 and I.11). Even though the upper 450 m of the section were not recovered, the 3365 m deep ARI-1 well provided a more complete and reliable record for Oligocene and Miocene interval than the NMA-1 well. Early Eocene sediments, however, were not encountered in the ARI-1 well. Sidewall core and cuttings samples were analyzed to determine the lithology, biostratigraphy and paleobathymetry of the sediments (ter Keurs, 1991; Brian O’Neill, pers. com.). A biostratigraphic study was conducted on 44 foraminifer samples (Fig. I.12), 10 nannoplankton samples and 10 palynomorph samples. The well logging in ARI-1 included gamma ray, sonic, resistivity and density logs. In addition, a vertical seismic profiling (VSP) survey was conducted. The results of the VSP survey provided reliable
Figure I.10. Correlation between the seismic and well data for ARI-1 well. Segment of depth-converted pre-stack migrated Shell seismic line E470 and its correlation with the lithological and biostratigraphic log of ARI-1 well.
Figure I.11. Location of Shell ARI-1 well on the segment of seismic line E470 (see Figure I.7 for line location). The well was drilled to test the seismically defined dip-closed structure. The well's biostratigraphy and gamma ray log are shown. The seismic section was split to show location of the well.
Figure I.12. Biostratigraphy and paleobathymetry of ARI-1 well based on the analysis of foraminifera (Brian O'Neill, pers. com.).

Bathymetric zones represent the following water depth ranges:

- **inner neritic**: 0-20 m
- **middle neritic**: 20-100 m
- **outer neritic**: 100-200 m
- **upper bathyal**: 200-500 m
- **middle bathyal**: 500-1000 m
- **lower bathyal**: 1000-2000 m
- **abyssal**: >2000 m

The left end of horizontal bars in the paleobathymetry column represents the weighted average of upper depth limits based on all species occurring in the sample. The right end of the horizontal bar represents basinward correction for suboptimal range bias, or "best estimate" of relative water depth. The tick mark represents the deepest upper-depth limit recorded for any species in the sample.

The percent planktonics curve shows the percentage of planktonic foraminifera relative to total foraminifera (planktonics and benthics) plotted sample-by-sample. Dark grey color denotes percent benthics and light grey pattern denotes percent planktonics.

Cosine theta (Davis, 1973) and Otsuka (Romesburg, 1984) are dissimilarity coefficients and are designed to classify stratigraphic boundaries by indicating paleontological "breaks" in data. Adjacent samples with similar fossil assemblages will have cosine theta and Otsuka values near 100 and adjacent samples with dissimilar assemblages will have cosine theta and Otsuka values noticeably less than 100.
### ARI-1

#### DEPTH b.s.f., m

<table>
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<tr>
<th>DEPTH b.s.f., m</th>
<th>BIO STRATIGRAPHY</th>
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<tbody>
<tr>
<td>500</td>
<td>N16</td>
</tr>
<tr>
<td>1000</td>
<td>N15, N14, N13, N10 to N12, N7 to N8</td>
</tr>
<tr>
<td>1500</td>
<td>Upper N5 to N6</td>
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<tr>
<td>2000</td>
<td>Lower N5, N4, N3</td>
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<td>2500</td>
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#### PALEO BATHYMETRY

- Coastal neritic
- Inner neritic
- Mid neritic
- Outer neritic
- Abyssal
- Upper bathyal
- Mid bathyal
- Lower bathyal

#### NEW SPECIES (FORAMS)

- Sample 1
- Sample 2
- Sample 3

#### COSINE THETA

- Sample 1
- Sample 2
- Sample 3

#### OTSUKA OSTRACODE ABUNDANCE

- Sample 1
- Sample 2
- Sample 3

#### Percent Planktonics

- Sample 1
- Sample 2
- Sample 3

#### Sample Gap

- Major discontinuity
sonic velocities of carbonate sediments that enhanced depth conversion of seismic data (Fig. I.10).

Both ODP Sites 714 and 715 lay outside the Shell and Elf seismic data coverage (Fig. I.2). ODP Site 715 was drilled on the eastern margin of the Maldives Ridge in 2266 m of water. The objective of this site was to penetrate the basaltic basement underlying the carbonate sediments. The well drilled through 211 m of carbonate sediments and continued for another 76.6 m into the volcanic basement. The upper 104 m of carbonate sediments consisted of foraminifera-bearing nannofossil oozes and chalks of the early (?) Miocene to late Pleistocene ages (Backman et al., 1988). The 107 m section between the basalts and pelagic oozes was made of shallow-water late early Eocene carbonates showing a deepening-upward sequence. A late early Eocene age was assigned to this unit based upon the benthic foraminifera assemblage (Nicora and Premoli Silva, 1990). The transition between the basalts and shallow-water limestones was recovered, and because the limestone beds were interbedded with the weathered basalt flows, the age of the volcanic eruptions could have been 3 to 4 Ma younger than the $^{40}Ar/^{39}Ar$ age. The lower part of the unit was a grainstone with solitary and colonial coral fragments and bryozoans indicating shallow reefal environment (Nicora and Premoli Silva, 1990). The section above this unit was represented by packstones containing large intact benthic foraminifera, bryozoans, gastropods and mollusks, implying a shallow-water shelf or platform environment. The upper part of the unit was wackestone with benthic
foraminifera, fragments of brachiopods, pelecypods, and bivalves, interpreted as deposits from a slightly deeper environment than the underlying sediments (Backman et al., 1988).

The objective of ODP Site 714, drilled in 2038 m of water in the proximity of the Site 715 in 2038 m of water, was to retrieve a complete late Neogene sequence of peri-platform aragonite-bearing oozes (Figure I.2). 233 m of Pleistocene to late Oligocene sediments were recovered showing a continuous record of sedimentation, with the exception of a hiatus from the late Pleistocene to the late Miocene in the upper section (0.5-8 Ma, Backman et al., 1988). The Miocene sediments contained extremely well preserved foraminifera and calcareous nannofossils. The upper unit of the Miocene-late Oligocene sediments (20-120 m) consisted of foraminifera-bearing nannofossil oozes with a minor amount of clay. The lower unit (120-233 m) was composed of moderately burrowed clay-bearing, foraminifera-nannofossil chalk.

The ODP Site 716 well was drilled in 544 m of water in the northern part of the Inner Sea in the proximity of the Shell seismic profiles and may be projected onto the Shell line E110 (Fig. I.13). The drilling recovered 262 m of foraminifera-bearing nannofossil oozes, grading into chalk downhole (Backman et al., 1988). The section contains Pleistocene to late Miocene sediments without any hiatuses (Fig. I.13), thus complementing the Inner Sea Shell ARI-1 well where the upper section was not recovered.
Figure 1.13. Location and stratigraphy of ODP Site 716 projected on Shell seismic line E100. See line location on Figure 1.7.
Litho- and biostratigraphic framework

_Eocene_

Three wells (NMA-1, ARI-1, and ODP Site 715) recovered the basaltic basement-sediment transition in the Maldives and provided somewhat conflicting information on the age, lithology, and depositional environment of the earliest carbonate sediments deposited in the Maldives. ODP Site 715 recovered late-early Eocene shallow water carbonate facies with abundant solitary and colonial corals, bryozoans, and large benthic foraminifera (Nicora and Premoli Silva, 1990). This suggests that carbonate production in the Maldives was initiated as early as in the late early Eocene in shallow-water environments.

The oldest sediments recovered in the Elf NMA-1 well consisted of inner-neritic dolomitized limestones with large benthic foraminifera (Fig. I.8). The age of the sediments directly overlying the basement was determined to be early Eocene by Lehmann and Pons (1986) based upon the large benthic foraminifera *Nummulites, Miliolidae, Alveolinidae*, and *Buliminidae*, the presence of which indicates deposition in shallow-water shelf environments (Hallock and Glenn, 1986).

The lowermost sediments overlying the basaltic basement in ARI-1 well were thin calcareous dark grey-reddish shales of the latest Eocene age (Fig. I.10). The depositional environments of these shales was determined to be coastal-shallow-marine (ter Keurs,
Coastal conditions were inferred from the well-preserved ostracod assemblages and occasional land plant material (coal). Marginal to fairly good Type III source rocks related to plant material were detected in a sample from 3228 m (ter Keurs, 1991). The palynological assemblages from this interval contained pollen of plants *Nypa fruticans*, *Oncosperma*, palms, *Avicennia*, *Aglaya*, *Santiria*, *Rhizopora* and *Palaquium* (ter Keurs, 1991). The overlying limestones contained abundant *Miliolidae* indicating shallow-marine, possibly restricted lagoon conditions (Hallock and Glenn, 1986).

Aubert and Droxlé (1996) suggested three alternatives to explain the age and lithofacies discrepancy between the NMA-1 and ARI-1 wells oldest carbonate sediments overlying the volcanic basement. The first explanation may be due to an error in the biostratigraphic analysis; the second suggests a more recent episode of volcanism in the proximity of the ARI-1 well that concealed the early Eocene sediments; and the third follows the rationale of Purdy and Bertram (1993) who proposed a “doming scenario”. Under this scenario, the mantle plume responsible for the origin of the volcanic ridge had a pronounced domal topography with its flanks submerged in water, while the central part was exposed. The thermal cooling than caused the collapse and subsequent structural subsidence of the central part of the dome leading to accumulation of thick sedimentary section that lacks the early Eocene sediments in the central part of the ridge where ARI-1 well was drilled.
Early Oligocene

In the NMA-1 well, sediments of early Oligocene age were determined in the interval from 1493 to 1432 m (Lehmann and Pons, 1986), and a 300-m section below was not recovered (Fig. I.8). The presence of *Nummulites* and other large foraminifera indicates that the deposition of sediments occurred in an inner shelf environment. In contrast, the lower Oligocene section in ARI-1 well (Fig. I.10) is much thicker (1053 m, interval 2220-3273 m), being composed of recrystallized limestones with few dolostone beds, rare red shales and traces of fine sand, possibly of volcanic origin (ter Keurs, 1991). The reworking of Eocene sediments is deduced from the occurrence of large foraminifer *Alveolina* in the lower part of the interval. More detailed chronostratigraphic subdivision on the lower Oligocene section is hampered by sample gaps in this interval (Fig. I.12). Planktonic foraminifera were essentially absent in this interval (Fig. I.12) and other microfossils were poorly preserved. Based on the diagnostic fauna, inner-neritic to middle-neritic depositional environment was determined for this interval.

Late Oligocene

A 10-m thick layer (2135-2145 m) of reddish and gray shales interbedded with marly limestone and containing minor amounts of pyrite and mica overlying the early Oligocene section was recovered in ARI-1 well (Fig. I.10). The age of the unit has been defined within the NP24 nannofossil zone (early late Oligocene age). The deposition of these shales is interpreted as being due to an abrupt shallowing event, based on the well’s
paleobathymetry (ter Keurs, 1991). The data indicates a shift from middle-outer neritic to inner neritic-coastal environments based upon the large increase in ostracoda abundance (Fig. I.12), the drop of the pelagic count, and the presence of large shallow water foraminifera such as miliolids (*Quinqueloculina spp.*) and rotalides (*Ammonia beccarii*), whose habitat is limited to the photic zone (Hallock and Glenn, 1986). On the gamma ray log, this interval corresponds to a strong positive kick (Fig. I.10).

The interval between 1895 and 2135 m is composed of recrystallized limestones interbedded with dolostones. The gamma ray log for this depth shows a variable pattern, with blocky low values pattern suggesting a “clean” limestone in the lower part, becoming progressively variable in the upper section indicating an increased shale content. Inner-neritic to middle-neritic depositional environment, with initial deepening followed by upward shallowing has been determined for the late Oligocene section in ARI-1 (ter Keurs, 1991).

In the NMA-1 well, the boundary between Oligocene and early Miocene is not defined and the 1188-1402 m interval is assigned the late Oligocene-early Miocene age (Lehmann and Pons, 1986).

*Early Miocene*

The early Miocene interval in ARI-1 well is 405 m thick (1887-1482 m) and is composed of fine white crystalline limestones with occasional bioclasts and intercalations of shales. Type I/II immature algal source rocks with high organic carbon content were
reported from the 1790-2020 m interval. These rocks with high TOC values were represented by dark grey-black shale intercalations.

The early Miocene section in ARI-1 well has been divided into two chronostratigraphic sections based on the planktonic foraminifera assemblages: Aquitanian (1846-1887 m) and Burdigalian (1482-1846 m). Inner-middle neritic to outer neritic depositional environment was determined for the early Miocene sediments of ARI-1. The mix of shallow marine (larger foraminifera) and deeper marine plankton/benthos assemblages suggest partial reworking of shallow marine sediments.

In the NMA-1 well, the early Miocene part of the section is represented by pelagic biomicrites with laminated black calcareous shales.

**Middle Miocene**

The middle Miocene sediments comprise the 1145-1482 m interval in the ARI-1 well. The sediments were fine crystalline limestones deposited in the middle/outer neritic-bathyal environment. In the NMA-1 well, the middle Miocene sediments (884-1128 m) were deposited in the outer neritic-upper bathyal environment (Lehmann and Pons, 1986).

**Late Miocene**

Only the lower part of the late Miocene sediments was recovered in ARI-well (805-1145 m). The sediments comprised marly planktonic ooze thought to be deposited in the upper bathyal environment based on the presence of the deep marine calcareous and arenaceous benthonics and abundant planktonic foraminifera. Foraminifera-bearing
 calcareous oozes of late Miocene age were also recovered at the ODP Site 716. The late Miocene sediments were not recovered in the NMA-1 well.

**Plio-Pleistocene**

No Plio-Pleistocene sediments were recovered from the ARI-1 well. In the NMA-1 well, the Plio-Pleistocene age was determined for the interval between 201 and 347 m, the top 150 m of sediments not being recovered. The Plio-Pleistocene sediments comprised inner neritic limestones with corals, algae, pelecypods, bryozoans and foraminifera (Lehmann and Pons, 1986).

Pleistocene sediments obtained from the ODP Site 716 cores were represented by periplatform calcareous oozes with abundant foraminifera. These sediments contained significant amount of shallow carbonate bank-derived aragonite needles (Droxler et al., 1990; Malone et al., 1990).

**Seismic data**

*Shell seismic data*

Shell data set consists of a 6000-km regular grid of 2-D seismic lines (Fig. 1.7). All lines are located within the Inner Sea covering a 275 by 50 km area. The grid consists of 88 E-W lines, 18 N-S lines and one NE-SW line. The data were collected in 1989 by GECO and were processed at the CGG center in Paris. The 60-fold seismic data were collected using four air gun arrays with individual 4804 in$^3$ gun capacity, towed at a depth
of 6 m and recorded with a 3000-m long 240-channel streamer with 12.5-m group spacing. The recording length was 6 seconds. Data processing included spherical divergence and geometrical spreading compensation, filtering, predictive deconvolution, 2-km spaced velocity analysis, NMO correction, phase compensation, and time migration. The quality of the data is good to excellent, in particular in the upper 2 seconds. Imaging problems do occur in the deeper part of the section under the carbonate reefal margins, sometimes impairing the platform-to-basin correlation. Dense data coverage, however, creates an opportunity to loop-correlate the reflectors to solve some of the imaging problems.

Resolution of seismic data

The dominant frequency of the seismic data was sampled on different seismic sections. The average dominant frequency at 1 second two-way travel time (TWT) was 50 Hz, 35 Hz at 1.5 s, and 25 Hz at 2 s. The vertical resolution of the zero-phase wavelet is estimated as ¼ of the wavelength (Kallweit and Wood, 1982). The vertical resolution of the Shell seismic data is therefore 10 m at 1 s TWT, 15 m at 1.5 s TWT, and 25 m at 2 s of TWT.

Seismic modeling of carbonate rock outcrops by several workers has demonstrated that the imaging of true stratigraphic geometries is highly dependent on the resolution of the seismic data. For example, seismic modeling of carbonate exposures in Dolomitic Alps, Italy (Stafleu and Schlager, 1993), Vercors, south-east France (Stafleu et
al., 1994), and Montagna della Maiella in Italy (Anselmetti et al., 1997), led to the conclusion that the modeled outcrop geometries were not portrayed correctly by the seismic data with a 25-50 Hz frequency bandwidth (the standard frequency range for industry and also the frequency of the seismic data from the Maldives). Moreover, misleading pseudo-unconformities have been reported from seismic sections modeled with the 25-50 Hz frequency bandwidth (Stafleu and Schlager, 1993; Stafleu et al., 1994). At the higher frequencies such as 60 Hz, however, all of the modeled supersequences are usually recognized (e.g. Anselmetti et al., 1997).

The difference between the outcrop modeling and the carbonate sediments in the Maldives is the seismic velocity of the material. The measured seismic velocities of the outcropping carbonate rocks ranged from 3.1 to 6.5 km/s, with the majority of the lithofacies characterized by seismic velocities between 4.5 and 6 km/s. In contrast, the seismic interval velocities of the commonly poorly lithified, and relatively uncompacted (in the upper section) carbonate mudstones and grainstones in the Maldives range from 2.0 to 5.7 km/s, but most commonly stay within a 2.0 to 3.6 km/s interval. This has been based on the results of the vertical seismic profile (VSP) in the ARI-1 well and stacking velocities. This means that the actual vertical resolution of the seismic profiles in the Maldives at the 25-50 Hz frequency is at least twice as high as the resolution of the modeled outcrops with the same frequency band. It is therefore concluded that the major sequences of this study are imaged correctly on the seismic sections. Seismic resolution does pose a limitation on the ability to recognize stational geometries on seismic profiles, but
resolution is sufficient in the seismic data presented in this study to recognize sedimentary sequences thicker than 10 m and their relationships in the upper part of the seismic sections in the Maldives.

Comparison of Shell and Elf seismic data sets

The more recent Shell seismic data used in this study has three advantages over the older Elf seismic data interpreted by Aubert and Droxler (1992, 1996) and Purdy and Bertram (1993). First, the overall quality of the Shell seismic data is far superior because of its higher resolution and more advanced processing. Multiple diffractions caused by dipping events such as platform margins and reefs impaired the interpretation of dipping events on the mixed phase, unmigrated Elf data. Second, all Shell seismic data was acquired and processed during one campaign. The Elf data set, however, consists of six different vintages of data, each with different acquisition and processing parameters, making the correlation between some of the lines difficult. The third advantage lies in having a dense, regularly spaced grid with 2-4 km line spacing. This provides a better three-dimensional image of platform geometries and allows for a better understanding of section variations along strike. The Shell data set, however, is limited to the Inner Sea, and utilization of the Elf seismic lines collected in the atoll lagoons is necessary to complete the picture of the entire carbonate system evolution. This is complicated by the fact that the data quality of the Elf seismic collected inside the atolls is poor due to uneven bathymetry and ‘ringing’ from the hard sandy lagoon floor.
Methods and correlation of seismic and well data

Seismic stratigraphy

The core data set used in this study is the 6000-km grid of 2-D multi-channel seismic profiles acquired by Royal Dutch Shell. Seismic interpretation of essential seismic lines was first conducted on paper sections. The sections were interpreted using the standard approach based on the evaluation of seismic facies, reflection terminations and geometries, and seismic attributes. Once the entire data set was available in digital format, the data were loaded on a UNIX workstation at Shell Bellaire Technology Center in Houston, Texas. Seismic units, their bounding horizons, and basement faults were correlated across the entire grid using Landmark’s SeisWorks 2D™ interpretation software. Seismically incoherent zones were commonly present under the carbonate margin buildups, making shelf-to-basin correlation difficult on some of the seismic sections. Nevertheless, loop-correlation around the incoherent areas was possible due to the high density of the seismic grid. The horizon and fault picks were exported into Landmark’s Zmap+™ mapping package and a series of time-structure and isopach maps were created using the software’s algorithms.

Ties between the seismic and wells

The correlation of the Shell ARI-1 well to seismic data was accomplished using the results of the vertical seismic profile (VSP) survey. The synthetic seismogram based
on the velocities from the survey showed excellent correlation with the seismic section, indicating good velocity control. The segment of the seismic line E470 on which the ARI-1 well was drilled (Fig. I.14), was pre-stack depth migrated and depth-converted (Fig. I.10). The correlation between the ARI-1 well and the seismic data is crucial for the definition of chrono- and lithostratigraphic packages.

The results of the check shot survey for the Elf NMA-1 well were not available. The NMA-1 well was tied to a seismic profile by using the velocities from the edited sonic log (Fig. I.9). ODP Site 716 was not logged and no velocity information was available. In order to tie the well to a proximal seismic line, a simple depth-time conversion was performed using stacking velocities and the velocities expected from the interpreted seismic facies (Fig. I.13).

*Sequence Stratigraphy*

The sedimentary section of the Maldives carbonate platform was divided into twelve major stratigraphic sequences based on seismic interpretation, lithology and age information from the wells (Table I.1). Although of variable thickness and duration, the sequences are characterized by distinct stratal geometries and seismic attributes. Each
Table I.1. Sequence and horizons used in the seismic interpretation of the Maldive sections.

<table>
<thead>
<tr>
<th>Sequence name</th>
<th>Sequence age</th>
<th>Upper Bounding Horizon</th>
<th>Sequence Scale</th>
</tr>
</thead>
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<td>late Paleocene</td>
<td>volcanic basement</td>
<td>regional</td>
</tr>
<tr>
<td>Eoc</td>
<td>Early Eocene</td>
<td>E1</td>
<td>regional</td>
</tr>
<tr>
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<td>Early early Oligocene</td>
<td>E-O-1</td>
<td>regional</td>
</tr>
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<td>E-Olig 2</td>
<td>Early late Oligocene</td>
<td>NP24</td>
<td>regional</td>
</tr>
<tr>
<td>L-Olig</td>
<td>Late Oligocene</td>
<td>O/M</td>
<td>regional</td>
</tr>
<tr>
<td>E-Mio 1</td>
<td>Early early Miocene</td>
<td>EM1</td>
<td>regional</td>
</tr>
<tr>
<td>E-Mio 2</td>
<td>Early late Miocene</td>
<td>E/MM</td>
<td>regional</td>
</tr>
<tr>
<td>M1</td>
<td>Middle Miocene</td>
<td>MM1</td>
<td>local</td>
</tr>
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<td>local</td>
</tr>
<tr>
<td>M3</td>
<td>Middle Miocene</td>
<td>MM3</td>
<td>regional</td>
</tr>
<tr>
<td>M4</td>
<td>Middle Miocene</td>
<td>MM4</td>
<td>local</td>
</tr>
<tr>
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<td>MM5</td>
<td>regional</td>
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<td>L-Mio 1</td>
<td>Early late Miocene</td>
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<td>regional</td>
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THE VOLCANIC BASEMENT OF THE MALDIVES

Nature and Timing of Volcanic Emplacement

The nature and age of the Maldivian basement is based on the ODP Site 715 and the two industry exploration wells, NMA-1 and ARI-1. Site 715, located on the eastern oceanward slope of the Maldivian archipelago, recovered 76.6 m of basalt lava flows. Three limestone units with basalt clasts were interbedded with the basalt lava flows in the upper part of this section. The upper parts of the individual basalt flows were weathered and oxidized, and this has been interpreted as the lava flows having been erupted under subaerial conditions (Backman et al., 1988). $^{40}\text{Ar}/^{39}\text{Ar}$ dating of these basalts yielded an age of 57.2±1.8 Ma (Duncan and Hargraves, 1990).

The Elf Aquitaine well NMA-1, drilled in the North Male atoll lagoon, penetrated 115.5 m of basalts below the carbonate section. Alternating hard and soft, weathered 10-m thick lava beds were composed of black to green aphanitic basalts with olivine. The age of these basalts was determined with the $^{40}\text{Ar}/^{39}\text{Ar}$ method to be 55 Ma (Aubert and Droxl er, 1996). This is consistent with the age from the ODP Site 715 and proves that these two dates provide reliable information about the timing of the basalt basement formation.
The Shell ARI-1 well was drilled in the central part of the Inner Sea (Fig. I.7). In its bottom section, the well encountered 50 m of weathered basalts. The severe degree of basalt alteration prevented reliable dating.

Structure of the Volcanic Basement

The basaltic basement underlying the carbonate edifice of the Maldivian carbonate platform is highly faulted. Normal faults are easily recognized on the seismic section and were reported by Aubert and Droxler (1992, 1996) and Purdy and Bertram (1993) from the Elf Aquitaine data set. Aubert and Droxler (1996) separated the volcanics underlying the carbonate sediments into the “acoustic basement” and “lava flow unit”. The “acoustic basement” was described as chaotic or transparent seismic facies cut by numerous faults. A strong reflector marked the upper boundary of this unit. The overlying “lava flow unit” was assigned to the late Paleocene age and was described as semi-parallel to chaotic reflectors of strong amplitude. According to Aubert (1994) and Aubert and Droxler (1996), many faults terminate within the ‘lava flow unit’ and do not extend into the overlying section.

Based on the Shell seismic data, the interpretation of the basement structure in this study was different, as an “acoustic basement” as described by Aubert and Droxler (1996) was not recognized. On the seismic sections tied to the wells, a strong reflector marks the top of the volcanic flows (Fig. I.14). The horizon picked as the high-amplitude
Figure I.14. Uninterpreted seismic line E470 and location of Shell ARI-1 well. See line location on Figure I.7.
reflector separating the basalts from the shallower carbonate section was assigned the name "volcanic basement" (Fig. I.15). The flat-lying parallel low-amplitude reflectors below were interpreted as noise and unsuppressed multiples. Although some reflectors below the top of the volcanics horizon may be primary, their seismic expression is weak and does not allow correlation across the basin. In contrast, the section overlying the volcanic basement is commonly represented by continuous parallel reflectors interpreted as shallow water carbonates.

Purdy and Bertram (1993) defined their sequence boundary #1 as the top of the volcanic basement unconformably overlain by the monotonous sequence of parallel high-amplitude reflectors. The "volcanic basement" horizon in this study is equivalent to the sequence boundary #1 of Purdy and Bertram (1993) and to the top of the "lava flow unit" of Aubert and Droxler (1996).

The volcanic basement is dissected by a series of normal faults forming an axial graben system (Fig. I.15). Fault-plane reflections are uncommon and the structural interpretation was based on identifying stratal terminations against the fault. The interpretation of fault planes was complicated by a lack of clear demarcation between the discontinuous strata of the hanging wall and footwall blocks. The faults appear listric on some seismic sections.

Two deep, relatively narrow grabens with a NNE-SSW orientation were mapped (Figs. I.16A and I.16B). The "southern graben" extends for at least a 100 km.
Figure I.15. Interpreted seismic line E470 and location of Shell ARI-1 well. See line location on Figure I.7. Refer to Table I.1 for horizon and sequence chart.
Figure I.16A. Time-structure map of the volcanic basement. Two deep graben systems have NNE-SSW orientation.
Its width varies from 4 to 8 km from south to north. The "northern" graben stretches for a minimum of 120 km. Its southern end is narrow but the graben bifurcates going north and is complicated by an upthrown central horst block in the northern part. The western "arm" of the northern graben becomes shallow at the northern extent of the Shell seismic grid while the eastern "arm" is prominent and likely to continue to the north toward the present day Fadifolu atoll.

The basement grabens were already described and mapped by Purdy and Bertram (1993), Aubert (1994), and Aubert and Droxler (1996) from the multi-channel seismic sections of Elf data set. The position and orientation of the grabens are similar to the interpretation presented herein; however, the interpretation in this study presents a much more refined and precise picture due to the density of seismic grid and superior seismic data quality. Purdy and Bertram (1993) extended the "northern graben" farther north between the present day atolls of Malosmadulu and Fadifolu and possibly under Tiladummati (their Fig. 25).

The symmetrical grabens are formed by major bounding faults, however, the fault pattern is commonly more complex, with generations of faults displacing the sediments within the grabens themselves. The overall throw of the faults varies from tens of meters up to 1800 m (e.g. Line E730, Fig. 1.17). The graben-bounding faults have dips of 31 to 47 degrees on the east-west depth-converted seismic sections. The lower values of the inclinations may be related to the oblique crossing by the seismic profile or a variation in the dip of the faults along strike. The fault blocks rarely show evidence of rotation
Figure I.17. Uninterpreted (A) and interpreted (B and C) Shell line E730.
although some of the blocks bear evidence of tilting which occurred long after the grabens were formed.

Topographically high basement blocks also display normal faulting with vertical displacement typically not exceeding 300 m. These faults do not appear to be spatially continuous and can be rarely correlated on more than two seismic lines. The orientation of these faults is generally parallel to the orientation of the graben-bounding faults, with the exception of a horst block bounded by SW-NE faults east of North Nilandu atoll (Fig. I.16).

A structural trend with a SW-NE orientation is observed on the basement time-structure map (Fig. I.16). This structure trend is not expressed on seismic profiles in the form of faults. This trend may be related to deeper basement faults, possibly of strike slip origin.

**Mechanism of Basement Fault Formation**

Aubert and Droxler (1992, 1996), Purdy and Bertram (1993), and Aubert (1994) described the basement fault distribution as an en-echelon graben system. According to their interpretation, left-lateral strike-slip movement formed this pattern. Purdy and Bertram (1993) suggested that the grabens were the near-surface expression of the left-lateral transform fault called the Chagos fracture zone (McKenzie and Sclater, 1971).
Aubert and Droxler (1996) related the basin deformation to a phase of divergent wrench tectonics, possibly associated with the initial stages of the India-Asia collision.

The interpretation of the Shell seismic data does not provide any clear indication that the graben systems are pull-apart structures linking major strike-slip faults as suggested by Aubert and Droxler (1992) and Aubert (1994). It is possible that major strike-slip faults are located outside of the Shell data grid; however, unambiguous strike-slip faults have not been reported from the Elf data set. Purdy and Bertram (1993) related the origin of the graben system to the Chagos transform fracture zone. Transform faults are strike-slip faults that appear as very steep (vertical) faults on seismic sections. Although the graben faults of the Maldives appear subvertical on vertically exaggerated seismic profiles, their true dips are too shallow (maximum 47 degrees) to imply a strike-slip component. There is also little evidence for the “en-echelon” structures” as the graben faults appear to be continuous and can be correlated for great distances. It is also important to note here that there is no indication of a fault connecting the two grabens oriented normal to the graben’s axis. The southern end of the northern graben seems to shallow up and “die out” east of the central part of Ari atoll; and the southern graben possibly ends southwest of the present South Male atoll. An extensional regime with active crustal-extension is probably responsible for the formation of the described structures, however, the exact mechanism responsible for such an extension is not clear. Returning to the hypothesis of Sheth (1999), the Laccadive-Chagos ridge may not be an expression of a stationary deep mantle plume under a moving plate but a “southerly
propagating fracture tapping a shallow, enriched mantle of the Indian Ocean” or a “huge shear zone…” that “reflects the stress field of the Indian Oceanic lithosphere” (p. 22).

Timing of the Fault Movement

The key to unraveling the mechanism of fault formation lies within the relative timing of structures. However, estimation of the precise timing of the initiation and cessation of the extension responsible for the graben formations in the Maldives is not easy. The onset of the extension is difficult to pin-point because no reliable reflectors that could be correlated exist within the basement fault blocks. Assuming that there were no hiatus between the eruption of the basalt lavas and the deposition of the overlying carbonate sediments, the minimum age of the onset is the age of the top of the basaltic lava flows (dated in the NMA-1 well and ODP Site 715 as late Paleocene, 55-57 Ma).

The examination of the stratal geometries of the sediments filling the grabens may provide the information on the end of the extension. The parallel reflectors of sediment fill within grabens imply uniform horizontal subsidence, and divergent reflectors imply a rotational motion of the hanging wall block (Cartwright, 1991). In the Maldives, the footwall blocks commonly bear evidence of erosional truncation, and fault scarps are commonly present on seismic profiles (Fig. I.17). Chaotic and prograding seismic facies interpreted as fans composed of eroded footwall block material are commonly recognized at the margins of the graben fill (Fig. I.18). The end of the extension is marked by parallel
Figure 1.18. Interpreted segment of line E130 showing a cross section through the northern graben. Note the fault scarp fans (shaded) interfering with graben fill deposits represented by continuous parallel reflectors.
onlap onto the fault scarp if the relief between the hanging wall and the footwall blocks was compensated, or by a marker bed with uniform thickness across the upper tip of the fault. The interpretation of Shell seismic shows that movement along the graben faults appears to have ceased sometime during the late early Oligocene time. From the late Paleocene to the early Oligocene, the grabens served as deeper seaways and were filled mainly with material shed from the surrounding higher structural blocks. Since the late early Oligocene, the graben faults remained dormant. Few minor faults are seen in the shallower section, but those that do exist are caused by the differential sediment compaction across the graben margins and not major tectonic movements.

The tectonic regime in the area of the Maldives may be characterized today as quiescent due to their location within the rigid Indian Plate (Fig. 1.3). This was not true for the late Paleocene-late early Oligocene time interval. Crustal stretching and possibly rifting affected the volcanic basement and the overlying section. The basaltic basement was faulted and sediments, with time, filled the negative structures and smoothed the initial relief. This is in contrast to the model of Aubert and Droxler (1996) who suggested that the basalt lava flows significantly smoothed the faulted “acoustic basement” and “created a flat topography that promoted the onset of the Maldive carbonate system” (p. 503).
SEISMIC STRATIGRAPHY OF THE MALDIVE CARBONATE SYSTEM

Results of Previous Studies Based upon the Elf Data Set

The general evolution of the Maldive platform was studied by Aubert and Droxler (1992, 1996), Purdy and Bertram (1993), and Aubert (1994) on the basis of the Elf Aquitaine seismic data set collected in early 1970s (Fig. 1.6). Despite the quality of the Elf seismic data and gaps in the spatial coverage, the main features and sequences of the Maldive platform were correctly recognized by these previous workers.

Purdy and Bertram (1993), from selected Elf seismic lines, defined three major sequence boundaries in the Maldives and interpreted them as erosional unconformities related to subaerial exposures. Their sequence boundary #1 was picked at the unconformity between the volcanics and overlying nummulitic carbonates. Sequence boundary #2 was characterized by the subtle erosional truncation of underlying reflectors and onlap of overlying reflectors. Based on the seismic tie to the NMA-1 well, sequence boundary #2 of Purdy and Bertram (1993) separates the early Oligocene shallow-water carbonates from the outer neritic-upper bathyal pelagic carbonates of the late Oligocene or early Miocene age. The pelagic carbonates graded upwards into prograding sequences. Sequence boundary #3 was characterized by considerable relief. It was overlain by a sequence whose upper surface paralleled the atoll’s present day bathymetry. This
unconformity was interpreted by Purdy and Bertram (1993) as the base of the Pliocene-Pleistocene section.

Aubert (1994) and Aubert and Droxler (1996) defined ten seismic units defined by a seismic stratigraphic analysis, calibrated by NMA-1 and ARI-1 wells. The correlation of regional unconformities using the Elf seismic was hampered by uncollapsed diffractions off dipping surfaces due to the unmigrated nature of the data and an incoherent seismic zone below the carbonate margin buildups. The interpretation of the two lower units, the "acoustic basement" and "lava flow unit", was discussed earlier. The overlying sedimentary section was divided into eight units. The lowest sedimentary seismic unit is Paleogene 1 (P1). It is represented by low-amplitude parallel reflectors in the basinal troughs, and high-amplitude layered to mounded reflectors in the shelf areas. The unit was assigned an Eocene-early Oligocene age. Paleogene 2 (P2) unit, of late Oligocene age, comprises basin-fill sediments and high-amplitude reflectors overlying the exposed early Oligocene shelf. The P2 unit is capped by a strong-amplitude reflector that was interpreted as the latest Oligocene regional drowning unconformity. Neogene 1 (N1) pelagic unit of Aubert and Droxler (1996) unit showed complete infilling of the troughs and burial of the preexisting platforms. The N1 reflectors onlap the Oligocene-Miocene drowning unconformity. In some locations, pelagic facies are replaced in the updip direction by "mounds". The age of the N1 unit is early Miocene to early middle Miocene.

The Neogene 2 (N2) unit is characterized by a basin-wide progradation towards the Inner Sea. Neogene units N3, N4 and N5 (late Miocene to early Pliocene) also show
progradation along some margins, usually underneath the present day atolls, and also fill the central seaway. Localized drowning and channel erosion is observed in the peripheral part of the basin and explains the segmentation of some banks. The late Pliocene-Pleistocene (PP) aggradational unit does not display progradation and is associated with the transition from flat-top carbonate banks to the irregular atoll morphologies seen today. A basinward shift observed at the base of this unit was interpreted as a response to significant lowering of sea level and exposure of the carbonate shelves. The fall was dated at ~3.0-2.4 Ma and corresponds to the onset of the Northern Hemisphere glaciation.

Seismic Stratigraphy: This Study

In this study, the stratigraphic section is subdivided into twelve main seismic sequences based on the seismic stratigraphic analysis of the Shell data set (Table I.1, Figs. I.15 and I.19-I.21). A more detailed seismic interpretation was conducted locally within the prograding sequences where a more refined interpretation was possible. The correlation of seismic units across the shelf-to-basin transition was difficult in the deeper part of the section due to the rapid lateral seismic facies change.

An alphanumerical system was used to assign the names to the horizons and the sequences they bound (Table I.1). Seismic units were correlated to the wells and their age was determined from biostratigraphic control (Fig. I.10).
Fig. I.19. Interpreted seismic line E130. See Figure I.7 for line location.
Figure I.20. Interpreted seismic line E520. See Figure I.7 for line location.
Figure I.21. Interpreted N-S seismic line N100. See Figure I.7 for line location.
Description of Seismic Facies and Stratal Geometries

Eocene Sequence (Eoc)

The lowest sequence Eoc rests unconformably on the faulted volcanic basement. The Eoc reflectors commonly onlap onto the upper surface of the footwall blocks and show basin-fill onlap within the grabens. The onlap is particularly pronounced in the east-central part of the basin where the basement block east of the graben is tilted to the east (Fig. I.17). The Eocene section here thins in this area to the west towards the graben fault.

The Eocene sequence has a variable thickness and is characterized by significant lateral variations of seismic facies, in particular, between the section deposited on top of the elevated blocks and the basin facies confined to the grabens. The seismic facies vary from high-amplitude parallel continuous to discontinuous mounded reflectors.

The top of the Eoc unit is a strong reflector picked as horizon E1. Although the resolution and quality of the seismic data is the poorest in this deeper portion due to the seismic wave attenuation and imaging problems under the carbonate margins, a number of lines provide good images of the banks with defined slopes and shelf breaks (Fig. I.22). Seismic facies interpreted as neritic carbonates overlay the volcanic basement and are established on the flat tops of the tectonic blocks that were topographically high. Seismic facies comprising the platform interior are typically strong-amplitude parallel reflectors and the facies forming the margin and slope are more mounded and discontinuous.
Figure I.22. Uninterpreted (A) and interpreted (B) segment of line E400 showing: Eocene bank with clearly defined bank margin; early Oligocene backstepping and aggrading bank, rimmed late Oligocene bank with lagoon and patch reefs.
Within the grabens, the seismic facies of the Eocene sequence may be separated into two general types: continuous parallel or divergent reflectors of moderate amplitude, and discontinuous to chaotic reflectors locally forming mounds that thicken away from the fault planes. The former are interpreted as a combination of peri-platform and pelagic sediments and the latter are thought to represent fans composed of the fault scarp material (Fig. 1.18).

*Early Oligocene Sequence (E-Olig 1)*

The age of this sequence is defined as early-early Oligocene but cannot be determined more precisely due to the significant sample gaps and abundant reworked material in the ARI-1 well. Within the sequence, the differentiation between platform and basin depositional settings is significant. Similar to the Eoc sequence, the platform facies are found on top of the structurally high tectonic blocks, and the basin facies are confined to the fault-controlled graben features. The typical platform seismic facies of the E-Olig 1 sequence are high-amplitude continuous reflectors overlying the Eocene platform sediments. Locally, the base reflectors of E-Olig 1 sequence onlap on the E1 horizon. The E-Olig reflectors commonly continue for great distances across any particular seismic section, invoking in mind the popular “railroad tracks” metaphor. The reflectors forming the overlying sequence are also continuous and parallel but have lower seismic amplitude. This contrast is clearly visible on the seismic sections (Fig. 1.22) and serves as the main
diagnostic criterion for the definition of the horizon E-O-1 that forms the upper boundary of the E-Olig 1 sequence.

The platform reflectors of the E-Olig 1 sequence show significant vertical aggradation and possibly backstepping (Fig. I.22). However, local erosion surfaces are common and may be misinterpreted as backstepping. The correlative basinal seismic facies within the grabens differ from the platform facies. The basinal facies are typically made of low-amplitude parallel to divergent reflectors, with wedge-shaped discontinuous facies characteristic of fault-scarp fan deposits. On some seismic profiles, the E-Olig 1 basinal reflectors onlap onto the slope of the Eocene carbonate platform (Fig. I.22).

*Late Early Oligocene Sequence (E-Olig 2)*

A prominent high-amplitude reflector that can be correlated across the basin marks the top of sequence E-Olig 2. In ARI-1 well, it is clearly dated as of early late Oligocene age within the nannofossil zone NP24. The reflector shows local erosional truncation of the underlying seismic reflectors. This high-amplitude reflector was picked as horizon NP24 and it forms the upper boundary of the E-Olig 2 sequence.

As mentioned earlier, the reflectors of E-Olig 2 sequence have markedly lower amplitude than those of the underlying sequence E-Olig 1. The E-Olig 2 platform reflectors are also continuous and parallel. Local erosion surfaces within this sequence are observed but cannot be easily mapped due to the insufficient spacing of the seismic grid. The platform areas show simple vertical stacking. The differentiation between the
platform and basin facies is not as pronounced as in the deeper sequences because many of the pre-existing basins were substantially filled with sediments.

Many normal faults rooted in the basement disappear within this sequence, indicating a cessation of the movement along the graben faults by the end of the early Oligocene time.

*Late Oligocene Sequence (L-Olig)*

The depositional profile of sequence L-Olig differs from the previous E-Olig 2 sequence. Individual carbonate banks, separated by seaways whose position is largely controlled by the location of the underlying grabens, begin transformation from aggrading flat-top carbonate banks into banks with a defined marginal rim (Fig. I.23). The elevated rims are made of chaotic mounded reflectors.

The reflector that forms the top of the late Oligocene sequence has an irregular topography and is easily distinguished on all seismic sections (horizon O/M, Fig. I.23). The horizon O/M is equivalent to the boundary between the P2 and N1 units of Aubert and Droxler (1996) described as “the only major unconformity that can be easily identified in most areas [on the Elf seismic profiles]” (p. 510).

Seismic facies of the late Oligocene sequence are represented by the high-amplitude parallel continuous reflectors in the bank interior, mounded to chaotic reflectors forming the marginal rim buildup, and high-amplitude parallel reflectors representing the basin fill facies (Fig. I.23). This sequence is characterized by rapid
Figure I.23. Interpreted segment of line E370 through the late Oligocene Gaha-Male bank with an elevated rim and protected lagoon with numerous patch reefs. In the early Miocene, the rim of the bank continues to aggrade while the lagoon patch reefs drown and are buried with carbonate mud, expressed on seismic as parallel onlapping reflectors. See line location of Figure I.7.
changes that include the development of raised bank rims and the widespread establishment of patch reefs in the platform interior. Bank rims typically nucleate on the shelf break of the early late Oligocene banks. The rims rise above the bank interior and can have a relief up to a 100 m (Fig. I.23). The basinward slopes of the platform rim become progressively steeper and reach angles of 15 to 18 degrees. The platform facies also show backstepping and aggradation in the latest Oligocene. The parallel continuous reflectors of the platform interior give way to pronounced mounded reflectors separated by parallel ones (Fig. I.23). The overlying lower-amplitude parallel reflectors of the shallower early Miocene unit onlap on the mound flanks and form onlap fill in areas between the mounds. I interpret the mounded features as patch reefs growing in the lagoon. The size of the individual buildups varies and may be sometimes erroneously measured from an oblique 2-D cross section but the typical dimensions of the mounds are 500 m in diameter and up to 100 m high.

*Early-Early Miocene Sequence (E-Mio 1)*

The depositional signature of the E-Mio 1 sequence varies in different parts of the basin. In the east-central part, sequence E-Mio-1 is composed of low-amplitude parallel reflectors, marking a change in the depositional geometry from a rimmed platform with mounded patch reefs in the lagoon to typical "empty bucket" geometry (Kendall and Schlager, 1981; Schlager, 1981, 1992). The rim of the large central bank continues to build up vertically while the patch reefs in the lagoon cease to accrete and drown (Fig.
I.23). The low-amplitude parallel reflectors onlapping the mounds represent soft carbonate mud that buried the drowned patch reefs. Not all of the reefs in the lagoon drowned, and a few, in exceptional cases, continued to accrete. Where possible, the bank margin build-ups continued to backstep “leaping” to higher topographic elevations on the bank (Fig. I.24). The deposition within the seaways represents significant basinal aggradation and fill. The basin reflectors onlapping the carbonate bank slopes have high continuity and low amplitude.

*Late-Early Miocene Sequence (E-Mio 2)*

Horizon EM forms the base of sequence E-Mio 2 whereas horizon E/MM marks the top. The E-Mio 2 internal reflectors are parallel and continuous with low- to medium amplitudes. Locally, lense-shaped areas of chaotic reflectors are present. The rim of the central large carbonate bank is represented by chaotic mounded facies and continues to accrete vertically.

In the peripheral parts of the system, the flat top banks were established due to backstepping and migration of carbonate build-ups to higher areas. The banks exhibit vertical aggradation. Unfortunately, the edges of the banks are rarely imaged on the Shell seismic lines, as they are located either directly below or in the proximity of the modern atolls that served as a navigation hazard for the seismic vessel. Nevertheless, the edges of the flat-top banks are imaged on a few Shell seismic lines (Figs. I.24 and I.25), and
Fig. 1.24. Late Oligocene - early Miocene backstepping of the carbonate bank margins and the establishment of the late early Miocene flat-top banks. Interpreted segment of line E120.
Figure I.25. Interpreted Shell seismic line E120. Two "arms" of the northern graben cut the volcanic basement. The grabens are filled with both eroded and pelagic material. Late Oligocene-early Miocene banks back the Oligocene. Backstepping resulted in formation of aggrading flat-top carbonate banks on both sides. The banks...
volcanic basement. Eocene and early Oligocene neritic carbonates capped the high topographic blocks while the banks backstepped while the bank formed on the horst block in the center drowned at the end of the sides. The bank margins prograded from both east and west in the middle Miocene.
sufficient information on the location of other banks comes from the Elf data set (Purdy and Bertram, 1993; Aubert and Droxler, 1996).

**Middle Miocene Sequences (M1-M5)**

The middle Miocene sequences (M1-M5) are characterized by distinct progradation of the flat-top carbonate bank margins (Fig. I.25). The progradation is represented by clinoforms that thin out in the basinward direction where they are commonly represented by a single reflector or even fall below seismic resolution. Five individual prograding sequences are recognized based on the stratigraphic relationships, reflection terminations, and seismic facies. Two middle Miocene horizons (MM3 and MM5) were correlated regionally, and three (MM1, MM2 and MM4) just locally due to their limited spatial extend (Fig. I.26). Three large areas of progradation ("prograding complexes") were identified within the limits of the Shell grid. Each prograding sequence consists of two distinct packages: the lower strong-amplitude reflector package (SARP) that shows a basinward shift in onlap, and the upper weak-amplitude reflector package (WARP) that displays progressive onlap in the topset area, followed by toplap in the upper section of the package (Fig. I.26). Within the individual packages themselves, diagnostic geometries are commonly recognized. In the SARP of the sequence M4, progressive downstepping of the shelf break is clearly displayed (Fig. I.27). Sequence M5 is characterized by the convex-upward, "mounded" reflectors at the base of the sequence boundary at the toe of the slope (Fig. I.26). These convex-upward reflectors downlapping
Figure I.26. Interpreted segment of Line E130 showing five middle Miocene prograding sequences (numbered 1-5). Each sequence is divided into strong amplitude reflector packages (SARPs shaded areas) and weak amplitude reflector packages (WARPs, no color fill). Also shown are late Miocene sequences with onlapping wedges forming the lower part of the sequences.
Figure 1.27. Uninterpreted (A) and interpreted (B) segment of Line E130 showing three downstepping margins in the middle Miocene sequence M4. The downstepping produces a "forced regression" geometry which is diagnostic for falling relative sea level.
on the flat underlying reflector are omni-present in the basin and are unique to Sequence M5.

In the central part of the basin, and at the location of ARI-1 well, the reflectors that are the basin equivalent of the sequences M4 and M5, show a gradual transition from being continuous and parallel to highly discontinuous and sometimes chaotic (Fig. I.28). Aubert and Droser (1996) described this structure as resembling imbrication. The reflectors appear to be either displaced by subvertical faults or possibly showing steeply inclined foresets.

Late Miocene Sequences (L-Mio 1 and L-Mio 2)

Two late Miocene sequences were defined within the Maldivian carbonate system. The lower sequence (LM1) is characterized by very continuous, parallel low to medium amplitude reflectors (Fig. I.25), locally showing oblique low-angle progradation. The amplitude of the prograding reflectors is typically higher than of the parallel continuous reflectors. The lower part of the unit shows onlap on the underlying Sequence M5. In the proximity of Felidu atoll, a large slump is observed (Fig. I.29). The slump is represented by chaotic seismic reflectors that form a 5-km-long basinward-thickening wedge. The slump displays toe-thrust faults at its western end where the slumping deposits collided with the mounded debris-flow deposits of sequence M5. The lower surface of the slump appears to be truncating the underlying reflectors and is interpreted to be erosional.
Figure I.28. Interpreted segment of line E470 showing layer-bounded disrupted deposits of the middle Miocene sequences M4 and M5. Parallel continuous reflectors on the east grade into disrupted, "imbricated" reflectors to the west.
Figure 1.29. Large early late Miocene slump near Felidu atoll. Note imbricated toe-thrusts at the point where slump deposits hit the lobe-shaped gravity-flow deposits of the middle Miocene sequence M5. Also note erosive basal surface of the slump.
The upper late Miocene sequence L-Mio 2 is also composed of highly continuous parallel reflectors of medium amplitude resembling the proverbial "railroad tracks" reflectors. Locally, downcutting channels are present within the sequence. The width of the channels is up to 2 km wide, typically 500 m, and the depth is up to 100 m.

_Early Pliocene (EP) Sequence_

Early Pliocene and Late Pliocene-Pleistocene sequences on Shell seismic lines are composed of low-to-medium amplitude highly continuous reflectors, with small channels within the sequence. Low-angle oblique progradation is sometimes observed within the EP sequence. Shell seismic profiles are limited to the area of the Inner Sea, and on the Elf seismic lines acquired in the atoll lagoons the depositional signature of this sequence is obscured by the water-bottom multiples (Aubert and Drozer, 1996).

_Late Pliocene-Pleistocene Sequence (LP-P)_

The base of the LP-P sequence is a strong-amplitude reflector that is dated as base late Pliocene by correlation to the ODP Site 716 (Fig. I.13). Some of the Shell profiles go through the submerged banks and show oblique highly progradational nature of these sediments (Fig. I.26). According to Aubert and Drozer (1996), a major basinward shift in onlap is present at the base of this sequence. This shift is also observed on a few Shell seismic lines (Figs. I.9 and I.30).
Figure I.30. Interpreted segment of line E120 showing deep-water (>200 m) progradation of the carbonate bank margin north off Gaha Faro atoll and middle Miocene prograding sequences (M1-M5) of prograding Complex III, attached to the slope of the late early Miocene flat-top bank.
INTERPRETATION OF THE INTEGRATED SEISMIC AND WELL DATA

Five main stages in the evolution of the Maldives platform evolution are inferred based on a combined interpretation of the seismic and well information. Each stage of the platform development has been governed by specific factors that created certain stratal geometries. During the first part of the Maldives evolution (Stages 1 and 2), basement structure and regional tectonics played the major role in the distribution of the shallow carbonate platforms and deep seaways. The second part (Stages 3 through 5) is characterized by a quiescent tectonic regime, when the basement lows were significantly infilled and relative sea-level fluctuations played the dominant role in the development of the carbonate system.

Stage 1 (Eocene): establishment of shallow-water banks

The lowermost sequence Eoc represents the establishment of the carbonate platforms on top of the faulted volcanic basement. High-amplitude parallel reflectors of sequence Eoc onlapping the basement blocks are interpreted as shallow platform carbonate sediments. Based on the seismic interpretation and well information, it appears that shallow-water carbonate platforms with corals and large benthic foraminifera (Nicora and Premoli Silva, 1990) were established in the early, possibly late early Eocene, on top
of the uplifted tectonic blocks after the flooding of the basement. A shallow-water environment is inferred from the presence of *Nummulites* and other large benthic foraminifera that live in symbiosis with photosynthetic algae and are restricted to the shallow photic zone (Hallock and Glenn, 1986). The grabens served as deeper seaways devoid of shallow-water carbonate production but filled with the pelagic and peri-platform sediments in addition to the material shed from the adjacent areas of higher relief.

The interpretation of Shell seismic profiles in this study offers an alternative explanation of the age discrepancy between ARI-1 and NMA-1 lowermost sections. The horizon that marks the top of the Eocene platform (Eoc) has been correlated around the basin within the limits of the Shell grid and to the wells ARI-1 and NMA-1. The location of the ARI-1 well is on a “shoulder” attached to the graben and is topographically lower than the interpreted shallow water Eocene platform (Fig. I.15). It seems reasonable that most of the Eocene platform material simply bypassed the area where the well was drilled and was deposited within the graben. The reworked large benthic foraminifera and coals (plant fragments) seen in ARI-1 well are likely the remnant material brought in from the surrounding highs that was deposited on the “shoulder”.

The role of the reef-forming fauna in the establishment of the shallow platforms is not clear. Coral and bryozans like the ones recovered in Site 715 are capable of constructing a rigid framework (Nicora and Premoli Silva, 1990). Large benthic foraminifera such as *Nummulites* are also capable of producing significant accumulations,
at times described as “build-ups” (Heckel, 1974). Others (e.g. Aigner, 1982) consider the cementsed Nummulites-rich skeletal accumulations as platform-margin shoals rather than constructional buildups. It was not possible to positively identify any unambiguous carbonate “buildups” on the Shell seismic lines.

The succession recovered in Site 715 shows a shift from the coral-bryozoan reef environment into a shallow-water environment dominated by large benthic foraminifera (Nicora and Premoli Silva, 1990). It is possible that bryozoans and corals were responsible for creating rigid, wave-resistant buildups at the early stage of the Eocene platform development. The well-defined Eocene platform margin (Fig. 1.22) may have been formed by either such buildups, or may have been formed by marginal shoals.

The time-structure map of Horizon E1 (Fig. 1.31) displays the upper surface of the Eoc sequence. Although the margin of the Eocene platform can be defined only on a limited number of seismic lines, seismic interpretation shows that distribution of shallow-water carbonate facies is confined to the relatively flat surfaces on top of the high basement blocks. In other areas, platform margin appears to be positioned at or near the graben-bounding faults. This indicates that the establishment of the shallow-water production in the Eocene was strongly controlled by basement topography. Neritic carbonates were established on the high tectonic blocks and then aggraded and retreated (backstepped) in the Eocene, their configuration strongly mimicking the basement trends.
Figure 1.31. Time-structure map of horizon E1 (top Eoc sequence) showing two main NNE-SSW oriented grabens and areas of shallow carbonate sedimentation. Position of the Eocene bank margin observed on seismic data is also marked on the map. The position of the margin is difficult to establish in other places where it is thought to be at or the graben fault.
Movement along the graben faults continued in the Eocene. Deep seaways separating the platforms became partially filled with the material shed from the surrounding fault blocks, and by pelagic sediments.

Stage 2 (lower Oligocene – early late Oligocene): bank accretion and exposure

Carbonate banks that had become established on the faulted basement in the Eocene continued to aggrade in the early Oligocene. The system’s configuration was similar to that of the Eocene: carbonate platforms occupying the high topographic blocks and deep seaways existing in the fault-controlled graben areas. The continuous parallel reflectors of sequences E-Olig 1 and E-Olig 2 are interpreted as shallow water platform facies. Local erosional surfaces are commonly observed, however, the overall stratal pattern of the platform sections is aggradational, with some seismic sections showing backstepping. Basin-fill sediments are again likely to be a combination of pelagic sediments and the sediments shed from the surrounding highs. The time-structure map of the E-O 1 horizon (Fig. I.32) shows two NNE-trending fault-bounded deep basins that are connected by a shallow spillway.

Early Oligocene sediments recovered in NMA-1 well suggest that Nummulites and other larger foraminifera continued to play an important role in sediment production. The early Oligocene section in the ARI-1 well contains large amounts of reworked material. It
Figure I.32. Time-structure map of E-O-1 horizon showing two graben-controlled seaways and relatively flat high-relief areas of shallow marine sedimentation. The main graben faults are still active and show displacement.
is clear from the time-structure map of E-O 1 horizon (Fig. I.32) that ARI-1 was positioned in a semi-enclosed basin adjacent to the northern graben basin. This morphology explains the thick early Oligocene section in ARI-1. The mini-basin where the well was drilled apparently served as a trap for sediments shed off the adjacent high.

The vertical accretion of the platforms continued in the late early Oligocene. It is not clear what caused the change in the amplitude of the reflectors in the platform interior. The early-Oligocene platform interior facies were not penetrated by the ARI-1 well, and their partial recovery in NMA-1 does not provide data to adequately answer this question. The reflector strength is a result of an impedance contrast, with impedance being the product of the rock seismic velocity and density. Possible causes include changes in the carbonate-producing biota, rates of accretion, duration of the sedimentation cycles, and diagenetic alteration. As suggested by the NMA-1 well stratigraphic log (Fig. I.8), the high-amplitudes of the lowest Oligocene reflectors may represent interbedded limestones and dolomites, while the upper part of the lower Oligocene platform section is predominantly made up of limestones. Beds of dolostones were indeed found in the lower Oligocene section in both NMA-1 and ARI-1 wells. Dolomitization possibly occurred during the short-term lowstands that promoted the flux of meteoric waters.

The biostratigraphy and paleobathymetry data from the ARI-1 well (Fig. I.10) suggest that a significant sea level fall exposed the platform tops at the end of the early
Oligocene (nannofossil zone NP24). The strong reflector that forms the top of the E-Olig 2 sequence corresponds to a strong positive kick on the gamma ray log in ARI-1 well (Fig. I.10). In the well, the gamma ray response corresponds to the calcareous shales that were deposited under the coastal-inner neritic conditions. Although deposition of reworked shallow-water material cannot be ruled out completely, precise biostratigraphic dating of this interval and presence of shallow water microfossil assemblages strongly suggest a shallowing event. On seismic, the early Oligocene reflector that form the top of the E-Olig 2 sequence displays erosional truncation of the underlying reflectors, more commonly observed at the bank edges. Downstepping of the margins was not observed on the seismic sections. This implies a relatively short duration of exposure. In the bank interior, convex downward reflectors may indicate collapse zones related to karsting, but the interpretation of karst features on seismic profiles is commonly difficult (Purdy and Waltham, 1999).

The strong amplitude of the NP24 reflector may be related to exposure and formation of a hard cemented surface. If the platform exposure was indeed a relatively short-term event, the platform tops might have not been exposed long enough for significant seismic-scale collapse or karst features to develop. The time-structure map of the horizon NP24 (top of early Oligocene) shows the basin configuration immediately after the NP24 event (Fig. I.33). Two seaways whose shape and position still follow the basement graben trends are connected by a shallow passage. The depth of the seaways is less than in the Eocene and early Oligocene because they had been substantially filled.
Figure I.33. Time-structure map of NP24 horizon showing two seaways separating large carbonate banks: Gaha-Male, Ari-Horseburg, and Nilandu-Felidu. The two seaways are connected by a shallow spillway. The steep western margin of Gaha-Male is well illustrated by the closely-spaced contour lines.
with sediments. The carbonate platforms separated by the seaways have pronounced slopes and a relatively flat bank interior morphology. A large bank with a well-defined semi-continuous rim is delineated in the northeastern part of the Shell seismic grid (Fig. I.33) For the simplicity of reference, it was given a name “Gaha-Male” bank after the names of the modern atolls it is adjacent to. The outline of the western margin of Gaha-Male is particularly well defined (Fig. I.33), and the eastern margin continues outside of the Shell seismic grid under the present day atolls. The carbonate bank whose edge is present on the northern and northwestern part of the seismic grid received a name “Ari-Horseburg”, and the bank that occupies the southern part of the grid was named “Nilandu-Mulaku” (Fig. I.33).

The time-structure map (Fig. I.33) shows that ARI-1 well was drilled in a small, circular-shaped mini-basin, connected by shallow spillways to small basins to the north and south. The results of the paleowater depth analysis in ARI-1 showed that the abrupt shallowing and establishment of coastal-inner neritic conditions at the NP24 time (Fig. I.10). This means that the relative sea level fall must have exposed the top of the surrounding carbonate banks subjecting them to dissolution and karstification. The magnitude of this fall estimated solely on the paleobathymetry data is 20 to 100 m (a shift from middle-outer neritic to inner neritic-coastal environment). The magnitude of the relative sea level fall may also be estimated by taking the difference between the exposed bank top and the point in the basin where shallow conditions were established, assuming that the bank top was near or at sea level. The magnitude estimated from the seismic
appears to be larger (>200 m) than estimated from paleobathymetry in ARI-1. but the value derived from seismic is not corrected for post-depositional tilting, compaction, or water loading.

Stage 3 (latest Oligocene-early Miocene): development of rimmed banks, drowning of the lagoon; backstepping and aggradation of the margins

The carbonate banks apparently recovered from the sea level fall during the early-late Oligocene transition after the exposed bank tops were flooded and carbonate production was reestablished. The most significant change that occurred in the late Oligocene was the formation of the aggrading bank margin rims (Fig. I.23). The depositional bank profile changed from an aggraded shelf type to a rimmed bank with a protected lagoon. This change may be related to either change in biota, increased rate of relative sea level rise that forced the banks to build up vertically to keep up with the rising sea level, or a combination of the two. The time-structure map of the horizon O/M (top of L-Olig sequence, Figs. I.34 and I.35) illustrates the morphology of the rimmed banks and the elongated deep central basin. In particular, the outline of the western part of the Gaha-Male bank occupying the northeastern part of the Shell grid is clearly seen. The rim of the bank continues for tens of kilometers but is not present along the entire length of the platform (Figs. I.34-I.36). The bank part adjacent to the circular mini-basin where ARI-1 well is located lacks a raised rim and has a relatively gently slope. This
Figure I.34. Time-structure map of O/M horizon showing late Oligocene carbonate banks (Gaha-Male, Nilandu-Mulaku, and Ari-Horseburg), and the central seaway. The western margin of Gaha-Male bank is characterized by an elevated rim. Patch reefs are common in the Gaha-Male bank interior. A bank in the northern part (part of Ari-Horseburg large bank) drowned at the end of Oligocene. Banks also show bankstepping away from the central part.
Figure I.35. Elements of the late Oligocene carbonate system, prospective view of O/M horizon.
morphology may be explained by wind and current conditions that existed at the time. Raised marginal rims typically develop in the high-energy environments characterized by strong wind and currents. The area where the rim was not developed was possibly located in the “shadow zone” protected by the continuous western face of the bank.

Platform interior seismic facies also change vertically within the sequence. The flat-lying parallel reflectors are replaced by the mounded reflectors interpreted as lagoonal patch reefs (Fig. I.23). The establishment of the patch reefs is related to either an appearance of frame-building fauna or to a relative sea level rise that promoted the “keep up” of the bank interior, or a combination of the two.

The reflector that forms the top of the L-Olig sequence is equivalent to the boundary between the P2 and N1 units of Aubert and Droxler (1996). They interpreted this surface as a regional drowning unconformity suggesting a possible exposure and karstification of the platform prior to the drowning but stating that the “exposure was not necessarily responsible for the development of the pronounced topography” (p. 517). From the interpretation of the migrated Shell seismic data and detailed reconstruction of the bank geometries for this study, no evidence of exposure at the Oligocene-Miocene transition is seen. Interpretations show the drowning of the patch reef buildups in the platform lagoons being contemporaneous with the aggradation of the bank margin rims and the backstepping of the margin buildups that continues into the early Miocene.

Some of the platform interior buildups display abrupt facies differences between the buildup facies and the onlapping reflectors. It is expressed particularly well on a part
of the platform in the northern part of the seismic grid where the rim did not develop and the entire platform drowned (Fig. I.37). This bank was established on a horst block in the middle of the northern basement graben, and was partially surrounded by deeper waters (Figs. I.34-36). The surface that marks the top of the bank buildups does not appear to truncate the bank interior facies (Fig. I.37), however, some may argue that local erosion surfaces are present.

The surface that marks the top of the bank seismic facies is considered here to represent drowning without subaerial exposure. Local erosion surfaces commonly develop on carbonate platforms after drowning, when the platform rim does not protect the platform interior from the strong currents (Schlager, 1998). In the regional context, platform rim aggradation and backstepping indicate rising relative sea level. The subaerial exposure of the carbonate platforms is not required for carbonate platform drowning (e.g. Schlager, 1993, 1998).

The aggradation of the bank margin rims and backstepping of the carbonate margins continued in the early Miocene. The aggradation of the Gaha-Male bank rim is particularly robust (Fig. I.23) but the most significant change is the drowning of the patch reefs in the Gaha-Male lagoon. The resulting geometry is commonly termed “the empty bucket” (Kendall and Schlager, 1981; Schlager, 1981, 1992), wherein the growth potential of the platform rim allows it to keep up with the rising sea level but the growth potential of the platform interior is lower which leads to drowning. Low-amplitude parallel reflectors that onlap the patch reefs represent soft carbonate muds that partially
Figure 1.37. Drowning of a latest Oligocene bank, uninterpreted (A) and interpreted (B) segment of line E130. The vertical growth of the bank ends at the Oligocene-Miocene transition (horizon O/M), and the overlying parallel onlapping reflectors represent pelagic carbonates. The early Miocene drowning of this bank is synchronous with the backstepping of other bank margins. Small early Miocene buildups were possibly formed on the highs created by the morphology of the drowned bank.
fill the “bucket”. Few mounded reflectors are found in the platform interior in the section above the drowning surface (Fig. I.37). Those that are seen may represent either sediments draping the underlying reefal mounds, or actual younger mounds that nucleated on the topographic highs caused by differential compaction related to the underlying buildups. At the same time, the bank rim buildups all across the basin show dramatic backstepping. The buildups migrate updip onto higher elevation where possible. The gradient of many platforms is very gentle and the carbonate buildups “leap” laterally for as much as 8 km (Fig. I.24). The older buildups drown and stop accreting.

Significant infilling of the existing seaways characterizes the early part of the early Miocene. The material deposited in the basins is both pelagic and bank-derived. This is reflected in the sediments recovered in the ARI-1 well where the early Miocene section consists of both pelagic limestones with black organic-rich shales and platform-derived large benthic foraminifera. The volume of sediments deposited in the basins is significant and is difficult to explain by shedding from the backstepping banks alone. It seems realistic that the productivity of plankton increased at that time which was possibly related to an increase of nutrient supply.

The time-structure map for the top of the E-Mio 1 sequence (horizon EM1, Fig. I.38) is markedly different from the map of the O/M horizon (Fig. I.34) where the differentiation between the platforms and the central seaway is clear. Although the outline of the deeper central trough still resembles the shape of the latest Oligocene seaway, it is shallower than its predecessor. The map also shows the backstepped bank margins and
Figure 1.38. Time-structure map of horizon EM1 showing backstepping and partial drowning of the early Miocene banks. The rim of Gaha-Male bank continues to aggrade while bank interior drowns. The shape of the central seaway is still reminiscent of the basement graben structure. Ari-Horseburg bank backsteps away from the central seaway and Nilandu-Felidu bank separates into smaller banks.
the elevated rim of Gaha-Male bank. At this stage, the control of volcanic basement structure on the deposition and distribution of sediments has virtually vanished.

In summary, the depositional geometries of the early Miocene sequence strongly suggest a relative sea level rise of significant magnitude. The rates of this rise exceeded the growth potential of the platform interior and caused its drowning. The rapid rise caused the dramatic backstepping of the platform margins and accretion of the platform rims. Rapid sea level rise is also responsible for the deposition of the organic-rich shales identified as an immature algal source rock in ARI-1 well.

Vertical growth of the Gaha-Male bank rim terminated sometime in the late early Miocene. As evident from the seismic sections, the rim of Gaha-Male continued to accrete until the latest early Miocene (Fig. 1.23). The exact demise of the platform rim is difficult to establish due to the complicated lateral correlation between the chaotic rim seismic facies and the surrounding sediments. Other correlation difficulties include the variation of rim elevation along strike and seismic sections crossing the rim obliquely. On some of the seismic sections, the rim is draped by the middle Miocene sequences, while on other sections the overlying middle Miocene sediments show collapse directly above the rim. The Gaha-Male rim apparently kept up with the rising sea level in the early Miocene and drowned in the end of the early Miocene.

During the latest part of the early Miocene a series of carbonate banks were established on the periphery of the basin. Unfortunately, the majority of the banks are located under the present day atolls where the Shell grid does not extend. The edges of
some banks, however, are imaged on a few Shell seismic profiles (e.g. Fig. 1.39). Information on the location of other banks comes from the interpretation of the Elf data set (Aubert and Droxler, 1992, 1996; Purdy and Bertram, 1993). The banks were flat-topped and aggraded vertically. The amount of vertical aggradation at the end of the early Miocene is significant (200 ms TWT on Fig. 1.39). The aggradation of the flat-top banks on the periphery of the basin occurred in response to the relative sea level rise that began in the early late Oligocene and continued until the beginning of the middle Miocene.

**Stage 4 (Middle Miocene): Bank Margin Progradation**

The middle Miocene interval in the Maldives is characterized by the wide-spread progradation of the peripheral flat-top banks formed in the latest early Miocene. The prograding sequences are expressed on dip seismic profiles as clinoforms attached to the bank edges (Figs. 1.25, 1.26, and 1.39). The isopach map of the middle Miocene section based on the Shell seismic data set shows three ‘bulges’ representing the loci of significant sediment deposition (Fig. 1.40). The thickness increase is related to the accumulation of prograding sequences. The term “prograding complex” is used to describe the “laterally extensive and vertically significant set of two or more clinoforms” (Part II, this volume). Three prograding complexes are defined within the limits of the Inner Sea. Prograding complex I, located in the northwestern corner of the Shell grid, is best imaged by the set of Shell seismic lines. Other areas of progradation lie outside of
Figure 1.39. Aggradation of the latest early Miocene flat-top bank and the middle Miocene progradation on the A) uninterpreted and B) interpreted segments of line E120. Note downstepping geometry (forced regression) in sequence M2.
Figure I.40. Isopach map of the middle Miocene prograding sequences. Three prograding complexes (I, II, and III) were identified within the limits of the Shell grid. Middle Miocene progradation was also documented by Aubert (1994), Purdy and Bertram (1993), and Aubert and Droxl er (1996) on the Elf seismic lines collected in the atoll lagoons (shaded areas on the map). Middle Miocene was not present under Felidu atoll (Purdy and Bertram, 1993).
the Shell grid coverage according to the interpretation of the Elf data by Purdy and Bertram (1993) and Aubert and Droxler (1996).

The middle Miocene progradation was bi-directional (Aubert and Droxler, 1992; Purdy and Bertram, 1993) from east and west towards the central part of the basin. Purdy and Bertram (1993) also noted the absence of the middle Miocene progradation under the Felidu atoll. The interpretation of Shell data shows that the degree of progradation was more substantial from the western side.

Two distinct packages are identified within each prograding sequence (Fig. I.26). Strong-amplitude reflector packages (SARPs) are interpreted as sediments deposited during the falling relative sea level (Part II, this volume). This conclusion is drawn on the observation that SARPs show consistent downward shift in onlap and sometimes exhibit evidence of downstepping (Fig. I.27). Weak-amplitude reflector packages (WARPs) are interpreted as sediments deposited during a relative sea level rise and subsequent highstand. The lower part of a WARP topset shows progressive onlap, presumably coastal, and its upper part exhibits toplap attributed to the highstand progradation. Each of the prograding middle Miocene sequences, therefore, represents a complete relative sea level cycle. The difference in the seismic attributes between the SARPs and WARPs is related to their composition and diagenetic alteration (Part II, this volume).

The mounded seismic facies at the toe-of-slope in the middle Miocene prograding sequence M5 are interpreted as lobe-shaped gravity-flow deposits (Fig. I.26). Steepening
of the bank margins and a base level drop (Part II, this volume) likely caused the
deposition of these lobe-shaped debris-flow fans.

In the middle Miocene, the bank margins prograded on both sides towards the
central part of the basin for 8 to 10 km. Lateral outbuilding of bank margins was
facilitated by numerous small channels oriented normal to the bank edge (Part II, this
volume).

The entire middle Miocene section is interpreted as deposited in response to
relative sea level falls punctuated by flooding events. Relative sea level appears to have
been lower in the middle Miocene than at the end of the early Miocene, and the latest
early Miocene flat bank tops possibly remained subaerially exposed in the middle
Miocene. During the relative sea level lowstands, the carbonate factory shifted in the
basinward direction, and carbonate production was limited to a narrow zone on the bank
slope. Reworked material from the exposed bank parts was deposited downslope and
formed SARPs. During subsequent flooding, a larger part of the bank was flooded which
led to an increased amount of produced bank carbonate sediment.

In the central part of the basin, the basin counterpart of the top two middle
Miocene sequences (M4 and M5) displays the discontinuous seismic facies (Fig. I.28)
described as "resembling imbrication" by Aubert and Droxler (1996). These seismic
facies may represent a regionally extensive polygonal faults system related to volumetric
contraction during early dewatering (Cartwright and Dewhurst, 1998). The facies are
correlative to the downstepping within the sequence M4 and gravity flow deposits of
sequence M5. The formation of the laterally discontinuous basinal facies is possibly also related to a significant base level drop that caused the release of overpressure in these sediments (Part II, this volume).

The individual sequences between the three middle Miocene prograding complexes and the ARI-1 well were correlated on seismic, confirming that the deposition of individual prograding middle Miocene sequences in the set-apart prograding complexes was synchronous. This implies that the mechanism responsible for the formation of prograding sequences operated on the basin-wide scale and was not related to the local variations in subsidence.

The time-structure map of horizon MM1 (Fig. I.41) displays the topography of the basin immediately before the initiation of progradation. This map shows the edges of some of the late early Miocene elongated flat-top banks positioned on the periphery of the Maldives system, and a central trough.

The progradation of the flat-top bank margins was responsible for the infilling of the central trough. The time-structure map of horizon MM3 (Fig. I.42) shows the position of prograding bank margins that prograde towards the wide central seaway. Following this, the time-structure map of the horizon MM5 (Fig. I.43) shows the basin morphology at the end of the middle Miocene progradation. The edges of the prograding banks have advanced considerably towards the central shallow trough that is a predecessor of the present-day Inner Sea of the Maldives.
Figure 1.41. Time-structure map of horizon E/MM showing the central trough and margins of flat-top banks on its periphery.
Figure 1.42: Time-structure map of the middle Miocene horizon M03 showing a broad central trough and prograding flat-top bank margins.
Figure 1.43. Time-structure map of horizon MM5 showing prograding bank margins and the central trough.
Stage 5 (Late Miocene – Quaternary): basin fill, aggradation of flat-top banks and atolls

Interpretation of Shell seismic shows significant infilling of the central trough during the late Miocene time. The paleo-Inner Sea was partially filled with periplatform and pelagic sediments expressed as highly continuous parallel reflectors on seismic profiles (e.g. Fig. I.19). The exposed late early Miocene flat-top banks in the peripheral part of the system were flooded and began producing excess sediments, with large amounts shed from the bank tops. The bank margins also continued to prograde towards the central part of the basin. The progradation, however, was less significant than in the middle Miocene.

The shape of the Late Miocene prograding reflectors is commonly oblique versus the sigmoidal shape of the middle Miocene reflectors. The prograding geometry is commonly attributed to the energy of the depositional environment (Sangree and Widmier, 1977). Oblique prograding units are though to be characteristic for the high-energy environment whereas sigmoidal progradation is typical of low-energy environments with low sediment supply. This oblique progradation may occur in relatively deep water (>100 m). High-resolution seismic profiles from the Strait of Florida published by Mullins and Neumann (1979) show an example of deep-water (750 m) oblique progradation of a carbonate bank margin. This progradation on the western side
of the Bahamas is driven by deep-water currents resulting from the combination of the dominant wind patterns and sea floor topography.

The late Miocene slopes are characterized by a lower gradient than those of the middle Miocene. A large slump in the proximity of present-day Felidu atoll (Fig. I.29) suggests that at least some of the late Miocene bank margins were unstable. Unfortunately, Shell seismic lines do not continue under the Felidu atoll and the exact slope gradient of the failed margin is not available.

The time-structure map of the horizon LM1 (Fig. I.44) displays a shallow central trough and the edges of flat-top banks along its periphery. The late Miocene sediments recovered in the ARI-1 well and ODP Site 716 were composed predominantly of planktonic oozes.

Time-structure map of horizon LM/P (Fig. I.45) demonstrates further infilling of the paleo-Inner Sea trough and the edges of the prograding margins. The margins are represented as bulges west of the area between Horseburg and Ari atolls, south of Ari and east of North Nilandu atolls, and east of Gaha atoll. Other important features include channels that appear both between the prograding banks and within the sequences. The channels were formed in relatively deep water, their origin possibly related to strong bottom currents. Aubert and Droxler (1996) observed localized drowning and channel erosional events within their N3-N5 late Miocene-early Pliocene on the Elf seismic data.
Figure I.44. Time-structure map of horizon LM1 showing central trough and outbuilding bank margins.
Figure 1.45. Time-structure map of horizon LM/P showing central trough and outbuilding bank margins along its periphery. Also note the development of channel separating some of the banks.
During the early Pliocene, the basin aggradation continued, with channels being common within EP and LP-P sequences. The channels do not appear to be continuous and cannot be correlated for significant distances.

Horizon LP forms the base of the LP-P sequence (Figs. I.19-I.21). It was picked as a strong-amplitude reflector and is dated as base Late Pliocene from the correlation to the ODP Site 716 (Fig. I.13). This horizon is correlative to the base of the Late Pliocene-Pleistocene (PP) unit of Aubert and Droxler (1996). They reported that PP unit was associated with the regional transformation from flat-top carbonate banks to irregular platform morphologies. Aubert and Droxler (1996) documented a significant basinward shift in onlap at the base of the PP unit. This shift is also observed on the Shell seismic profiles (Fig. I.30). In addition, Purdy and Bertram (1993) and Aubert and Droxler (1996) reported collapse features under the atolls from the Elf seismic lines.

The time-structure map of the LP horizon (Fig. I.46) resembles the present day sea floor morphology (Fig. I.2). The shallow paleo-Inner Sea occupies the central part of the basin and gently sloping bank edges are present on both western side and eastern sides. Channels are present between some of the banks. Based on the composition of carbonate sediments from the ODP Site 716, the accumulation of carbonate material in the central part of the Inner Sea is mainly pelagic, with some flux of aragonite and high-Mg calcite bank-derived material (Droxler et al., 1990; Malone et al., 1990).
Shell seismic data set does not permit to address transition from the flat-top banks to atolls because the seismic lines do not continue inside the atolls lagoons. It is worth mentioning, however, that the present morphology of atolls is most commonly attributed to the periodic exposure and flooding of the flat-top banks (Purdy and Betram, 1993; Aubert and Droxler, 1996). Meteoric waters have been responsible for the dissolution of the carbonate material, creating a karstified topography that served as a substrate for the coral-algal organisms when the bank top was flooded again. This “antecedent karst theory” (Purdy and Betram, 1993) seems to fit with the significant fluctuations of sea level during the Pleistocene, related to the waxing and waning of the polar ice caps.

Another issue is the timing and mechanism of the formation of the channels separating the present day atolls, and the drowning of some banks. Aubert and Droxler (1996) emphasized the local character of drowning events that cannot be explained by differential subsidence due to the lack of recent tectonic activity and structural control. Purdy and Bertram (1993) suggested that environmental factors or the variable depth of the reflooded substrate were possible causes of localized drowning. Absence of the high-resolution shallow seismic data and cored drill holes on present atoll and drowned banks leaves a wide field for speculations on this matter.
TECTONIC CONTROL ON THE EVOLUTION OF THE MALDIVES

The tectonic subsidence in the Maldives is evaluated based on information from ARI-1 well. Subsidence modeling was performed using the BasinWorks software (PetroDynamics Software, Inc.). The program first plots the depth of each horizon after a correction for the compaction at their respective depths. The plot is then adjusted for paleobathymetry, showing the total subsidence for the bottom of the sediment column. Tectonic subsidence is derived by adjusting for isostatic compensation of the sediment load, assuming that the crust has no strength. Figure I.47 shows the total and tectonic subsidence for the ARI-1 well assuming a constant sea level. The plot shows a generally smooth subsidence with faster subsidence rates in the late Eocene - early Oligocene and lower rates for the rest of the time.

The reconstruction of the volcanic basement structure and carbonate platform development clearly shows that tectonic control was dominant during the early stages of the Maldives evolution. The distribution and configuration of the shallow carbonate platforms is intimately related to the systems of NNE-SSW grabens that displace the volcanic basement. Movement along the graben faults continued until the early Oligocene but its affect on sedimentation appears to have been progressively diminishing. Eocene and late Oligocene carbonate banks occupied the topographic basement highs while the graben areas were seaways devoid of shallow marine carbonate production. The grabens
Figure I.47. Total and tectonic subsidence curves for ARI-1 well. Positive value of subsidence at 25 Ma is possibly related to a biostratigraphic error or to an underestimated paleobathymetry value. Subsidence curves were built using BasinWorks software (PetroDynamics Software, Inc.)
became filled with the material exported from the bank tops combined with fault scarp fan deposits. Similar strong basement structural control on the distribution of carbonate platforms and banks has been long recognized (e.g. Fulthorpe and Schlanger, 1989; Cocozza and Gandin, 1990; Wilson, 1990; Bosence et al., 1998; Gomez-Perez et al., 1999).

In the Maldives, the tectonic basement structural trends are still recognizable on the Eocene and early Oligocene time-structure maps (Figs. I.31 and I.32). Carbonate banks that became established on localized topographic highs typically have steep margins and aggradational stratal geometry. The banks aggraded vertically during the periods of sea level transgression and highstand. The elevated platform rims commonly developed near the footwall block margins and have aggraded with time. During the periods of relative sea level lowstand, the bank tops became exposed, and platform slopes were subject to erosion, with the subsequent downslope redeposition of eroded material. Downstepping and progradation were not typical for the isolated carbonate platforms rooted on small-scale tectonic blocks, due to the high slope gradients and deep surrounding waters.

During the more recent part of the Maldives evolution, relative sea level has played a dominant role on the geographic distribution and development of carbonate banks. The cessation of tectonic activity at the end of the early Oligocene, combined with the transgression in the late Oligocene-early Miocene, drastically changed the configuration of the carbonate system. Dramatic backstepping of platform margins and
the infilling of the seaways muted the influence of the basement structure almost entirely. The middle Miocene progradation of carbonate banks further modified the morphology of the Maldive carbonate system.
MALDIVES OLIGOCENE/MIOCENE RELATIVE SEA LEVEL RECORD – IS THERE A GLOBAL COMPONENT?

Isolated carbonate platforms are commonly regarded as ‘dip sticks’ for past sea level changes (e.g. Ebertli and Ginsburg, 1989; Schlager, 1992). Carbonate production is confined to the photic zone and robust tropical platforms are highly sensitive to sea level fluctuations. Isolated carbonate platforms are detached from continental landmasses, and are therefore removed from large pulses of terrigenous sediments associated with tectonic uplifts or climate variation on land. The physiographic and tectonic setting of the Maldivian platform thus makes it a particularly attractive area to address the past sea level changes. The size of the Maldives platform is large enough that the stratal geometries similar to those on passive margins were developed. Located on the rigid part of the Indian plate and characterized by the quiescent tectonic regime for the last 30 Ma, the Maldives are subject to a relatively simple subsidence history related to the cooling of the oceanic crust, and water and sediment loading.

The relative sea level record established from stratal geometries and paleobathymetry information in the Maldives reflects the combined effect of the subsidence and eustatic sea level changes. Although the detailed reconstruction of sea level changes in the Maldives at this time is limited by the sparse well sampling, the overall evolution of this platform should be regarded in a global prospective. A more
precise analysis of the past eustatic changes recorded in the Maldive stratigraphy must await a comprehensive transect drilling campaign with substantial sediment recovery, much as attempted by the ODP drilling campaigns on the New Jersey Margin (Miller et al., 1996) and in the Bahamas (Eberli et al., 1997).

The results of the New Jersey Margin Project (Miller et al., 1996) and the Bahamas Transect (Eberli et al., 1997) confirmed that Neogene sea level fluctuations charted by the Exxon group in the 1970s and 1980s are global. Some discrepancies in timing exist, and a more precise correlation is precluded by the limits of biostratigraphic resolution. This argument forms the base for the critics of the global eustatic chart (e.g. Miall, 1992).

The other issue of great importance raised in sea-level studies is the absolute magnitudes of eustatic events. The knowledge of the magnitudes is crucial for the correlation of the events in different basins. An alternative, widely used proxy of past sea level for the 'ice-house world' is benthic foraminifera oxygen isotope data. The isotopic composition of sea water reflects the amount of continental ice due to the preferential sequestration of oxygen isotopes during evaporation. Calcareous shells secreted by marine fauna record the changes in the isotopic composition of sea water. Temperature and salinity of sea water also affect the oxygen isotope ratio. Bottom-dwelling fauna is less affected by the temperature and salinity variations in the water column.

Recent studies (Miller et al., 1996; Abreu and Haddad, 1999; Pekar and Miller, 1996) show that the timing of oxygen isotope events correlates with the major sequence
boundaries of the sea level chart of Haq et al. (1987). The estimates of the sea-level magnitudes derived from oxygen isotope data, however, are commonly much lower than magnitudes of eustatic events based on the measurements of coastal onlap (Pekar and Miller, 1996; Pekar et al., 1999).

The interpretation of the Shell seismic data presented in this study shows that the Eocene and early Oligocene evolution of carbonate platforms was strongly controlled by the basement topography, subsidence, and activity of the basement graben faults. This complex control, combined with a lower seismic resolution at this depth, and poor sediment sampling, does not make the differentiation between the subsidence and eustatic signals viable. Although some local erosion surfaces are recognized, the Eocene-early Oligocene Maldives platforms mainly aggraded and backstepped in response to the rising relative sea level, most likely primarily driven by basement subsidence. The late Oligocene and Miocene sediments, on the contrary, were deposited under a tectonically stable regime and are well characterized by the stratal geometries and paleobathymetry. The interpretations for the late Oligocene – middle Miocene section are next compared with the global sea level record established from the coastal onlap charts and oxygen isotope data.
Oxygen Isotope Record

The recent benthic foraminifera oxygen isotope data compilation by Zachos (1998) is one of the most complete records of oxygen isotope variations for the Neogene and late Paleogene (Fig. I.48). This compilation has well-recognizable trends. Eocene isotope values increase progressively throughout the epoch. The early Oligocene values are significantly heavier than the Eocene ones. The rapid increase in oxygen isotope values just above the Eocene/Oligocene boundary has been reported from a variety of deep-water sites. Miller et al. (1991) labeled the climax of this event as Oi-1. A detailed study of cores from the South Atlantic Ocean and southern Indian Ocean demonstrated that the early Oligocene event Oi-1 was a brief extreme cold interval (Zachos et al., 1996). The benthic foraminifera oxygen isotope values from these sites show a spectral peak at 41 kyr, suggesting control by the obliquity cycles. The event was related to the increased continental ice volume and a decline in deep-water temperatures. The $\delta^{18}O$ values rebound to the lighter values in the early Oligocene, staying just below 2‰. The early/late Oligocene transition is marked by the pronounced shift towards the heavier values with a subsequent return to the lighter values (Fig. I.48). Although a high-resolution record for this event is not currently available, a few early-late Oligocene data points have values that are more than 1‰ heavier than late early Oligocene values.
Figure I.48. Comparison between the benthic foraminifera oxygen isotope data (Zachos, 1998) and global sea level curve of Haq et al. (1987) for the Oligocene-middle Miocene time interval. The two records show similar trends: a drop of sea level in mid-Oligocene (~28.5 Ma), a sea level rise in the late Oligocene, and a sea level fall, composed of high-frequency events, in the middle Miocene.
Benthic foraminifera oxygen isotope values show a steady increase in the late Oligocene. The values are the lightest at the end of Oligocene (0.7%). The Oligocene-Miocene transition is characterized by a significant short-term increase in the isotope values when they become as heavy as 2‰. The values then quickly become lighter but remain 0.3‰ heavier than at the end of Oligocene. A high-resolution record of stable-isotope and percent coarse fraction data across the Oligocene/Miocene boundary was published by Zachos et al. (1997). The results of the densely sampled expanded nannofossil chalk section from the Hole 929A on Ceara rise in western equatorial Atlantic show a brief positive excursion of 1.2‰ (equivalent to Mi-1 event of Miller et al., 1991) just above the Oligocene/Miocene boundary. The 1.2‰ excursion was initiated 24.0 Ma and peaked at 23.7 Ma. The Mi-1 event, in detail, consists of high-frequency cycles of variable amplitude superimposed on a longer-term increase. The oxygen isotope record carries a strong response at the obliquity (40-kyr) period. This periodicity implies a high-latitude climate control (ice volume and/or temperature) on the oxygen isotope values at the Oligocene/Miocene transition. The 1.2‰ increase requires significant accumulation of continental ice, equal in volume to the present day East Antarctic Ice Sheet, or a bottom water temperature cooling of 5-6 °C, but was most likely related to a combination of the two. Dissolution was the major factor influencing the percent coarse fraction (%CF) (Zachos et al., 1997). The %CF is coupled with the δ¹⁸O in the Hole 924 record and this relationship may be explained by the flux of corrosive bottom waters
during the glacial periods. The even higher-resolution record from Site 926 (Ceara Rise) revealed large magnitude shifts in bottom water temperature and ice volume on a scale similar to events observed in the Quaternary (Zachos et al., 1999).

The late early Miocene oxygen isotope record has high-frequency oscillations but not significant magnitudes (Fig. I.48). The $\delta^{18}$O values become lighter at the end of the early Miocene and the middle Miocene record shows high-frequency variation superimposed on a long-term oxygen isotope increase. Wright and Miller (1993) compiled oxygen isotope data from three ODP sites in the North Atlantic and Southern Ocean. Their benthic oxygen isotope values show a long-term decrease from the earliest Miocene to the early middle Miocene, and a step-wise increase in the second part of the middle Miocene, beginning around 15 Ma. The values reach their maximum at the end of the middle Miocene.

Global Sea-Level Chart

The global sea level chart of Haq et al. (1987) is a modification of the coastal onlap chart (Vail et al., 1977) that was based on the compilation of records from seismic well, and outcrop data. In order to adequately compare the oxygen isotope records to the global sea level chart of Haq et al. (1987), the modified version of the chart adjusted to the modern geological time scale was used (Hardenbol et al., 1998; Berggren et al., 1995). The sea level chart shows a period of high sea level in the lower Oligocene. One of
the most prominent sea level falls is the mid-Oligocene sea level fall placed within the NP24 nannofossil biostratigraphic zone. The magnitude of this prominent drop at the Rupelian/Chattian boundary is 180 m on the global chart. This event, known as the “30 Ma fall”, is now dated at 28.5 Ma.

According to the sea-level chart (Haq et al., 1987), the upper Oligocene sea level rebounded slightly but remained significantly lower. In fact, the sea level never attained the same value that it had in the lower Oligocene. A sea-level rise occurred in the earliest Miocene time. The lower Miocene interval is characterized by five events of different magnitudes but none of them exceed 60 m. The middle Miocene shows a series of events superimposed on the broad sea-level fall that reaches its maximum at the middle/late Miocene transition. The magnitude of this event is 140 m but the cumulative amplitude of the step-wise middle Miocene sea-level fall is event greater. The late Miocene sea level shows a generally rising sea level punctuated by sea level falls of relative small amplitudes, with the exception of the Messinian event (90 m).

Maldives Sea Level Record in the Global Framework

The geologic history of the Maldives based on the interpretation of seismic and well data shows the shallowing of seaways and exposure of carbonate platforms during the early/late Oligocene transition (Fig. I.49A). In the ARI-1 well, the abrupt change from middle neritic to inner neritic-coastal environments was recorded and well-dated within
Figure I.49. Schematic evolution of stratal geometries of the Maldivian carbonate platform in the early/late Oligocene - middle Miocene.
the NP24 nannofossil zone. On seismic, downstepping was not seen and the NP24 event is expressed by a continuous high-amplitude reflector that locally displayed erosional truncation, with possible collapse features within bank interiors related to karsting. This event is synchronous with the “30 Ma” (28.5 Ma on corrected timescale) sea level fall on the Haq et al. (1987) chart and corresponds to the benthic foraminifera oxygen isotope increase on the compilation by Zachos (1998) (Fig. I.48).

The change in paleowater depth at the NP24 interval based on the ARI-1 paleobathymetry data alone may vary from 20 to 100 m. The magnitude measured from the seismic profile tied to ARI-1 yields a value as high as 275 m. This value is not corrected for water loading, compaction, or isostatic rebound, yet it appears to be unreasonably high. The overestimation of the fall magnitude may be due to the post-depositional tilting of the footwall block on top of which carbonate was established. In fact, the shallower early and late Miocene basin reflectors are dipping to the west indicating at least some post-Oligocene tilting. Despite the uncertainty about the magnitude of the mid-Oligocene fall, this shallowing event was significant, and the only one reliably recorded in the ARI-1 well log.

Abreu and Haddad (1999) suggested that the bulk rock oxygen isotopes from the Petrobras Well A provide an independent verification of the mid-Oligocene sea level fall as one of the largest in the Oligocene-early Miocene. On the other hand, other scientists, while not denying the existence of the mid-Oligocene fall, suggest much lower amplitudes for this event. The amplitude of the NP24 event estimated from a model using
combined benthic foraminifera biofacies and two-dimensional flexural backstripping results on the New Jersey Coastal Plain did not exceed 30 m (Pekar et al., 1999).

A glacio-eustatic mechanism is commonly considered responsible for creating the mid-Oligocene sea level fall observed in basins around the world, regardless of its amplitude. The eustatic fall is attributed to an increase in the amount of the continental ice in Antarctica. The late Oligocene time is characterized by the significant increase of ice-rafted material, and ice grounding events recorded by glacial erosion surfaces and tills that are thought to be related to the expansion of the East Antarctica Ice Sheet. Bartek et al. (1991) concluded that an extensive erosional event on the Ross Sea continental shelf was caused by the grounding of an expanded Antarctic ice sheet and indicated that the ice sheet reached the continental scale by the mid-Oligocene time. The intensity of the mid-Oligocene unconformity is particularly strong on the African continental slopes where this event is marked by widespread development of submarine canyons. In fact, the mid-Oligocene unconformity is the cornerstone of sequence stratigraphy because its appearance on seismic sections from offshore African basins was so obvious that it was the first to be recognized by the Exxon stratigraphy group in the 1970s (P. Vail, pers. com.). Burke (1996) pointed out that the mid-Oligocene unconformity may be expressed more strongly on the African continental margins than in other parts of the world. McGinnis et al. (1993) suggested that the magnitude of the mid-Oligocene sea level fall determined from stratal geometries was overestimated because of the flexural rebound along margins induced by the erosional unloading. The late Eocene and early Oligocene...
deep-water erosion was induced by currents related to the onset of the ice growth in Antarctica and establishment of pole to equator climatic gradients and cold water production. The removal of sediments due to erosion caused the isostatic rebound and accentuated the mid-Oligocene unconformity on narrow shelves (McGinnis et al., 1993). Burke (1996) argued that the mid-Oligocene glacial conditions persisted and therefore deep-water erosion should have continued as well. He suggested that the salt-rich deep water current flowing from Tethys into the Indian Ocean until about 16 Ma, as well as the establishment of the Benquela current, were responsible for the deep sea erosion that enhanced the mid-Oligocene unconformity around the African continent.

Hofmann et al. (1997) suggested that the mid-Oligocene cooling and the subsequent Antarctic ice growth were triggered by the eruption of massive Ethiopian flood basalts. Recent dating of the Ethiopian basalts placed the time of their eruption at about 30 Ma, with the duration of the eruption not exceeding 1 Ma. Hofmann et al. (1997) proposed that the injection of sulfur-rich aerosols and dust in the atmosphere related to the violent Ethiopian basalt eruptions accelerated the global cooling and aridity and ultimately enhanced the growth of continental ice.

The stratal geometries interpreted from the Maldives data demonstrate continuous aggradation and backstepping of the carbonate platform margins in the late Oligocene through the early Miocene (Fig. I.49B-D). Evidence is lacking for a significant sea-level fall at the Oligocene-Miocene transition as predicted by the benthic foraminifera oxygen isotope records (Miller et al., 1991, Zachos et al., 1997). Oxygen isotope data imply that
the early Miocene sea level was lower than during the latest Oligocene interglacial. Even if Mi-1 event is assumed to be short-term and the exposed platform tops were rapidly flooded, the carbonate platforms should have downstepped and developed at a lower topographic level than the late Miocene platforms.

On the contrary, the late Oligocene-early Miocene transition in the Maldives is marked by the switch from the rimmed Gaha-Male bank with lagoon patch reefs to the "empty bucket" morphology with the elevated rim and the drown lagoon (Fig. I.49C). This implies not sea level lowering but an increase in the rate of the sea level rise! The pronounced backstepping of the margin buildups also suggests accelerated rates of the sea level rise. The paleobathymetry data from the ARI-1 well also demonstrates an overall deepening of the basin (Figs. I.10 and I.12).

The discrepancy between the Maldives record and the oxygen isotope record in the early Miocene may be explained by significant regional subsidence in the Maldives, or by the temperature effect on the oxygen isotope data. The subsidence curve built on the ARI-1 data does not show accelerated subsidence rates in the late Oligocene-early Miocene (Fig. I.47), but reconstruction based on a single well in a large area such as the Maldives is not sufficient to rule it out. The temperature effect on the isotope data, however, appear to be a more plausible explanation of the discrepancy between the Maldives and isotope records in the light of the newly published study by Lear et al. (2000). The temperature-corrected benthic oxygen isotope record shows increasing isotope values in the late Oligocene and early Miocene (Fig. I.50). Based on the results of
Figure I.50. Temperature-corrected benthic foraminifera oxygen isotope data (from Lear et al., 2000).
the Mg/Ca-paleothermometry study, the decrease of the oxygen isotope values in the early Miocene was related to the cooling of deep water, not associated with the accumulation of continental ice (Lear et al., 2000). The sea level trend predicted from temperature-corrected benthic oxygen isotope record is in full accord with the Maldivian relative sea level record.

The reconstructed Maldivian sea-level record is also in general agreement with the global sea-level chart of Haq et al. (1987) showing a broad sea level rise in the late Oligocene-early Miocene punctuated by small-magnitude sea-level falls, of which the Oligocene/Miocene and Aquitanian/Burdigalian are more pronounced (Fig. 1.48).

The basin-wide synchronous progradation of the Maldivian bank margins in the middle Miocene is interpreted as a result of five complete sea-level cycles. The relative sea-level position in the middle Miocene appears to be lower than in the early Miocene. This interpretation is consistent with both benthic foraminifera oxygen isotope records and the Haq et al. (1987) sea level chart. Flower and Kennett (1994) stated that the middle Miocene time was the turnover point for the Cenozoic climate. The early part of the middle Miocene was characterized by major changes in ice volume, global sea level, and climate. During the second part, the East Antarctic Ice Sheet expanded significantly, causing major changes in deep water circulation and production. Abreu and Anderson (1998) argued for the initiation of the West Antarctica Ice Sheet in the early-middle Miocene and its repeated advances and retreats in the early stage. All lines of evidence
seem to suggest that significant sea-level falls and rises in the middle Miocene were caused by the expansions and reductions of the continental ice sheets in Antarctica.
ENVIRONMENTAL CHANGES AND SEDIMENT SUPPLY

Two general assumptions were made in the reconstruction of the Maldive platform evolution. First was the assumption that the geographic distribution of the biota at a given time in the vast Maldive carbonate platform was fairly uniform. The other assumption, related to the first one, was that the geographical variation of sediment supply within the areas of shallow-marine production was not significant. The role of both environmental-biological changes and sediment supply variations in creation of stratal patterns has been emphasized by some workers (e.g. Schlager, 1991, 1993). Although knowledge of these two variables in the Maldives is quite limited due to sparse sampling by drilling, their importance and potential influence on the sediment deposition in the Maldives will be addressed.

The environmental control on carbonate production is significant. It may result in shut-downs of the carbonate factory due to the nutrient poisoning or temperature stress, and ultimately lead to the drowning of the carbonate platform that is not sea-level related (e.g. Hallock and Schlager, 1986). The appearance of different types of frame-building biota through time may influence the type and morphology of the platform. The geographic variations in biota in the Maldives basin were probably not significant from Eocene to Late Miocene. Seismic interpretations show simultaneous aggradation, drowning, or progradation of individual platforms or banks which implies a basin-wide
control. This may not be true for the Plio-Pleistocene section. Aubert and Droxler (1996) emphasized the diachronous drowning of carbonate banks in the Plio-Pleistocene that was not related to rapid pulses in subsidence due to the absence of synsedimentary tectonic activity in the Neogene. Aubert and Droxler (1996) also attributed the local drowning to the development of channels separating the present day atolls. Purdy and Bertram (1993) ruled out the bank drowning due to the environmental differences on the grounds that temperature, salinity, nutrient supply and currents would not vary dramatically within the Maldives basin. They suggested that the depth of the substrate at the time of flooding caused the localized drowning of the banks. Ciarapica and Passeri (1993), based on their study of the modern Maldivian fauna, stress the importance of environmental control. The outbreaks of the vicious Indo-Pacific starfish, *Acanthaster planci*, capable of digesting coral polyps, are responsible for killing entire colonies of hermatypic corals. Other organisms (gastropods, asteroids, and echinoderms) are also capable of destroying large coral colonies which leads to accelerated reef erosion and, ultimately, their drowning.

Large benthic foraminifera played an important role in sediment production in the Eocene and Oligocene, when the Maldives area was part of the Tethyan province. Corals were reported from the ODP Site 715 (Nicora and Premoli Silva, 1990) but mounded seismic features identified as carbonate build-ups are nearly absent on the seismic sections in the Eocene and early Oligocene. The accretion of platform rims and the development of lagoon patch reefs in the late Oligocene is potentially related to the increasing role of the scleractinian corals in the production of rigid, wave-resistant
buildups. Coral reefs and associated buildups were established in a wide latitudinal range in the southeast Asia in the late Oligocene-early Miocene time (Fulthorpe and Schlanger, 1989). The closure of the Tethys Ocean in the Zagros Mountains in Iran occurred in the early Miocene at about 15 Ma. This closure helped to establish the modern conveyor-belt circulation in the oceans and separated the Mediterranean and Indo-Pacific provinces.

Another important implication on the carbonate sedimentation in the Maldives is the establishment of Asian monsoons. The elevation of the Himalayas and the Tibetan plateau is linked with the intensity of the monsoonal winds (Ruddiman and Kutzbach, 1991). The timing of the Asian monsoon inception is not well constrained. DeMenocal (1995), reporting results of the study focused on temporal variations of windblown dust concentrations in the deep-water cores, proposed that the Asian monsoon existed as early as 12 Ma ago. Other estimates are more conservative and place the first appearance of monsoons at 8 Ma ago (Kroon et al., 1991).

The wind-current regime in the Maldives is considered responsible for the bidirectional bank progradation towards the Inner Sea (Aubert and Droxlér, 1992, 1996; Purdy and Bertram, 1993). The question remains whether the monsoonal regime was responsible for the middle Miocene progradation, or the progradation was simply related to the filling of the available accommodation space in the sagging basin, as proposed by Purdy and Bertram (1993). The pronounced difference between the leeward and windward margin, related to the direction of the easterly trade winds, was demonstrated in the Bahamas both at present and in the subsurface (Eberli and Ginsburg, 1989). There,
the leeward margin of the Great Bahama Bank prograded for 25 km while the windward margin experienced near-vertical growth (Eberli and Ginsburg, 1989). The seasonal switch in monsoon wind direction may explain the bidirectional progradation in the Maldives in the middle Miocene. It is likely, however, that the wind and current patterns varied along the 800-km long Maldive platform. In the present-day Maldives archipelago, the wind and current direction and intensity varies geographically and, in addition, is modified by the local atoll topography (UNEP/IUCN, 1988). Interpretation of Shell seismic data shows that the middle Miocene progradation is better developed on the western side which may be related to the stronger westerly winds and currents at the time. Shell data, however, is limited to the Inner Sea, and this observation may be a result of aliasing.

The paleoproduction of the marine plankton introduces yet another uncertainty. The upwelling zones rich in nutrients cause plankton blooms responsible for deposition of thick pelagic sediments. The thick early Miocene infill of the seaways may be related to the position of the Maldives in a high-productivity zone.
CONCLUSIONS

Shallow marine carbonate banks were established in the Maldives in the Eocene on top of faulted volcanic basement. During the first part of the platform evolution, the basement structure controlled the geographic distribution of the shallow-water carbonate banks and their development. Two NNE-SSW trending grabens served as inter-platform seaways. The Eocene and early Oligocene signatures of the carbonate banks are mainly aggradational. The movement along the graben faults was continually diminishing and finally ceased in late early Oligocene.

During the second part of the Maldives evolution (late Oligocene-present), variations in relative sea level played a major role in shaping the platform development. A significant (20-100 m) short-term sea-level fall in the mid-Oligocene (~28.5 Ma) exposed the bank tops and resulted in a shallowing of seaways. The carbonate production resumed in the late Oligocene. A relative sea level rise in the late Oligocene-early Miocene is recorded in stratal geometries by the backstepping of bank margins and the development of platform rims. The Oligocene/Miocene transition is also marked by the drowning of the bank lagoons and creation of an “empty bucket” morphology that indicates accelerated rates of sea level rise.

The late-Oligocene-early Miocene flooding that resulted in significant backstepping and filling of the seaways muted basement structure control. Flat-top
carbonate banks were established in the peripheral parts of the basin at the end of the early Miocene. The banks aggraded significantly in the latest early Miocene. The bank margins prograded towards the center of the basin in the middle Miocene is response to five complete sea level cycles. Relative sea level in the middle Miocene remained below its early Miocene position, and bank tops were exposed. The bank tops were finally flooded in the late Miocene.

The late Miocene-early Pliocene period is characterized by basin infilling and accretion of carbonate flat-top banks on the periphery of the central trough. During Quaternary time, periodic exposure and flooding of carbonate banks resulted in transformation from a flat-top bank to atoll morphology.

The reconstructed relative sea-level history for the Maldive platform geometries shows good similarity with deep-water benthic oxygen isotope records in the Oligocene-middle Miocene. Although the magnitudes of sea level can not be precisely established from the Maldives data at this time, the general trends predicated from the isotope record find their verification in the depositional signature of the Maldives. The important discrepancy between the reconstructed Maldive record and oxygen record exists in the early Miocene. The Maldive data suggest a continuous sea level rise from the late Oligocene to early Miocene, while the isotope data implies a lowering of sea level in the earliest Miocene. This inconsistency disappears when the Maldive relative sea level record is compared with the temperature-corrected benthic foraminifera oxygen isotope
data which shows progressively lighter isotope values in the late Oligocene-early Miocene, indicating reduction of continental ice volume.
PART II.

CARBONATE BANK RESPONSE TO MIDDLE MIocene SEA LEVEL

FLUCTUATIONS IN THE MALDIVES (EQUATORIAL INDIan OCEAN)

ABSTRACT

The Maldives large isolated carbonate platform in the equatorial Indian Ocean is an ideal natural laboratory to study how carbonate platforms respond to relative sea level changes in the Miocene. The middle Miocene time in the Maldives is marked by pronounced bi-directional basin-wide progradation of carbonate bank margins towards the central part of the basin. The interpretation of a dense grid of Royal Dutch/Shell 2-D seismic data identified prograding depocenters and imaged the detailed internal architecture of the prograding sequences. Progradation of the carbonate bank margins started in the early middle Miocene. Three individual prograding complexes were located tens of km away from each other. Each prograding complex consists of five sequences bounded by unconformities and correlative conformities. Seismic interpretation shows that the deposition of sequences in individual prograding complexes was synchronous and was therefore driven by regional-scale mechanism.

One prograding complex, (complex I), is examined in detail in three dimensions. Each prograding sequence was subdivided into two packages: strong amplitude reflector packages (SARPs) and weak amplitude reflector packages (WARPs). SARPs consistently
display a basinward shift in onlap and are interpreted as falling system tract prograding wedges. WARPs are interpreted as having formed during the rise and highstand of relative sea level. Each sequence thus represents a complete sea-level cycle. Higher-order sea-level events can sometimes be recognized within the individual SARPs. The terminal middle Miocene sequence is characterized by voluminous downslope deposition of sediments interpreted as gravity flow deposits. This basin-wide event is interpreted as a response to a significant sea level fall at the end of the middle Miocene.

The magnitudes of relative sea level fluctuations are interpreted from the stratal patterns. The method used is based on the identification of “anchor points”, or diagnostic geometries that are direct indicators of relative sea level position. The interpretation suggests that prograding sequences have recorded a series of relative sea-level falls and rises of significant (50-100 m) magnitude. Middle Miocene relative sea level remained lower than sea level at the end of the early Miocene. The reconstructed relative sea level record shows good similarity with the global sea level record predicted from the benthic foraminifera oxygen isotope data. Various data suggest the episodic growth of the East Antarctica, and possibly West Antarctica ice sheets in the middle Miocene. The glacio-eustatic scenario is therefore invoked as the main driving mechanism for the middle Miocene progradation in the Maldives.
INTRODUCTION

Progradation is lateral outbuilding of sedimentary bodies and is common in various depositional settings (siliciclastic shelves, carbonate margins and banks, mixed silicilastics and carbonates). Prograding strata are easily recognized on both seismic and outcrop sections because of their characteristic clinoform shape in dip view. Progradation of carbonate platforms and bank margins has been described at different scales in the geological record of basins worldwide (e.g. Bosellini, 1984; Eberli and Ginsburg, 1989; Pomar, 1993; Sonnenfeld and Cross, 1993; Tinker, 1998). Prograding margins consist of vertically and laterally stacked unconformity-bounded sedimentary units or sequences. Each sequence represents a pulse in sedimentation controlled by interplay between sediment supply and accommodation space.

Prograding margins play an important role in basin evolution because they may rapidly change the geographic position and shape of a carbonate bank. Progradation can be responsible for the coalescence of isolated carbonate banks as in the Great Bahama Bank ("the Strait of Andros", Eberli and Ginsburg, 1989) and Saya de Malha Bank in the Indian Ocean, Purdy and Bertram (1993, their Figure 29).

In carbonate depositional systems, progradation has been documented for carbonate shelves (e.g. Tyrrell and Davis, 1989; Pomar, 1993), ramps (e.g. Sarg, 1988; Handford, 1995), and large isolated platforms (e.g. Eberli and Ginsburg, 1989). Progradation does not develop on small isolated carbonate platforms and atolls because
their steep slopes are surrounded by deep waters. In this case, progradation of the bank margins is prevented because the excess carbonate material produced on the platform tops is transported downslope by catastrophic events in a form of slumps and turbidite flows. As a result, the material shed from a small isolated platform or bank typically bypasses the slope and is deposited at its toe as sheets or aprons (e.g. Sarg, 1988; Handford and Loucks, 1993).

The interpretation of stratal geometries and stacking patterns of prograding sequences is commonly used to reconstruct their deposition and understand factors such as accommodation history and sediment supply. Other factors, such as compaction-related subsidence (Hunt et al., 1996; Hunt and Fitchen, 1999), environmental change (Schlager, 1993), and antecedent topography (e.g. Van Wagoner et al., 1990, Steckler et al., 1993; Owens et al., 1999) are considered by some scientists to play a major role in influencing sequence development. Reconstruction of the relative sea level record, based on geometrical patterns alone is hazardous, because similar geometries may represent different depositional features. In prograding strata, the offlap break (the clinoform rollover point) was considered to indicate the fairweather-wavebase at the depth of 5 to 20 m (Vail and Todd, 1981; Vail et al., 1984; Posamentier and Vail, 1988; Van Wagoner et al., 1988; Hunt and Tucker, 1993). Recent results of clinoform modeling in siliciclastic systems (Pirmez et al., 1998; Steckler et al., 1999) demonstrated that the rollover point may represent various water depths and is not diagnostic of the relict sea-level position.

Deep water (>200 m) sediments such as drift deposits may also prograde. High-resolution seismic profiles in the Straight of Florida on the northwest side of the Great
Bahama Bank and Little Bahama Bank show a deep-water (750 m) mound-like feature with oblique prograding internal reflectors (Mullins and Neumann, 1979; Mullins et al., 1980). This type of progradation is related to sediment transported by bottom currents, the Gulf Stream in the case of Florida Straight, and clearly is not controlled directly by changes in the accommodation space.

Many spectacular examples of prograding carbonate margins have been studied in outcrop and on seismic profiles. To this day, however, almost all of the progradation models, with a few exceptions (Osleger and Tinker, 1999), are based on 2-D cross sections. Because the deposition of sediments occurs in three-dimensional space, sediment distribution and the resulting stratigraphic geometries may vary significantly along strike. A single 2-D cross section is not likely to adequately represent the anatomy of the prograding margin, and a comprehensive 3-D model is desired for the documentation and full understanding of the process.

In the Maldives platform in the equatorial Indian Ocean, the middle Miocene prograding carbonate bank margins were identified by Aubert and Droxler (1992, 1996) and Purdy and Bertram (1993). Carbonate banks located at the periphery of a large elongated central basin prograded towards the center from both the east and west sides. These previous studies were based on the interpretation of unmigrated 2-D seismic profiles collected in the early 1970s by Elf Aquitaine. The reported progradation was unusual because: 1) it occurred on an isolated oceanic platform, and 2) it was bidirectional. Seismic quality and data coverage did not allow the detailed separation of the prograding sequences and interpretation of their geometries and stratigraphic patterns.
The data set used in this study consists of a grid of 2-D seismic lines collected in the Inner Sea of the Maldives during the 1989-1991 Royal Dutch/Shell exploration campaign. The dense seismic grid allows the detailed documentation of the internal architecture of the individual prograding bank margins. The along-strike variation of stratal geometries can be studied on seismic profiles of various spatial orientation. Constructing a series of isopach maps of individual prograding sequences gives a moving view on the evolution and spatial relationship of the clinoforms. The timing of the sequence formation is based on correlation to two deep industry wells (NMA-1 and ARI-1) and Ocean Drilling Program (ODP) Site 716.

The goals of this study were: 1) to document the morphology and internal architecture of prograding sequences in 3-D on a set of strike and dip lines; 2) to study the evolution of prograding margins through time; 3) to establish the timing of the progradation in the basin and determine the scale (local vs. regional) of the progradation in the Maldives; 4) to assess the mechanism and controls responsible for the formation of the prograding clinoforms; 5) to attempt the reconstruction of relative sea level (accommodation) history of the basin in the middle Miocene by interpreting stratal relationships and geometries; and 6) to address the reconstructed relative sea level history in a global sea level/climatic context by comparing it to the global sea-level record predicted from the benthic foraminifera oxygen isotope data.
GENERAL GEOLOGICAL SETTING OF THE MALDIVES

The Maldive Archipelago is located in the equatorial Indian Ocean and occupies the central part of the Chagos-Laccadives ridge. The atolls form an 800-km long chain extending in a north-south direction (Fig. II.1). The archipelago consists of 22 large atolls ranging in size from a few kilometers to tens of kilometers in diameter. The physiography of the Maldives is unique and makes the archipelago remarkably different from other isolated oceanic carbonate platforms and guyots. The first main difference is the large size of the Maldives platform. The second unique feature is two parallel chains of atolls in the central part of the archipelago separated by a 50-km wide relatively shallow (water depth 200-600 m) Inner Sea (Fig. II.2). The oceanward slopes of the atolls, in contrast, are steep and rapidly reach water depths in excess of 2500 m.

The aseismic Chagos-Laccadives ridge stretches for over 3000 km in a north-south direction from the southeast coast of India to south of the equator. The origin of the ridge is attributed to the Réunion hot spot activity (Duncan and Hargraves, 1990). The hot spot became active at around 65 Ma when the massive Deccan trap basalts erupted in India. The fixed position of the hot spot under the northward moving Indian Plate consequently created the Laccadives-Maldives-Chagos ridge, Saya de Malha Bank, Nazareth Bank and Mauritius. The present-day position of the hot spot is thought to be located under the island of Réunion (Duncan and Hargraves, 1990).
Figure II.1. Atolls of the Maldives Archipelago, central equatorial Indian Ocean. Black areas represent atoll islands, grey areas indicate atoll lagoons.
Figure II.2. Central Maldives atolls and bathymetry map. Locations of ODP Sites 714, 715 and 716 and industry wells ARI-1 and NMA-1 are also shown.
Previous studies in the Maldives based on the Elf-Aquitaine seismic data (Aubert and Droxler, 1992, 1996; Purdy and Bertram, 1993; Aubert, 1994) established the general evolution of the Maldives carbonate system. The Elf data set consisted of 6750 km of 2-D multi-channel seismic data, with 2700 km collected within the atoll lagoons. The seismic grid provided a good regional coverage. Wide grid spacing and marginal data quality permitted the identification of the main depositional environments and their evolution through time. Seismic processing did not include time migration and phase compensation, and diffractions from dipping events such as bank margins significantly distorted seismic displays and hampered interpretation. The lines acquired within the atoll lagoons were generally of poor quality due to the uneven bathymetry and the “ringing” effect from the hard bottom substratum of the lagoon floor.

The basalts underlying the carbonate edifice of the Maldives platform erupted subaerially at about 57 Ma (late Paleocene) based on the $^{40}\text{Ar}^{39}\text{Ar}$ dating of the samples recovered from the Leg 115 ODP Site 715 and the NMA-1 well (Duncan and Hargraves, 1990; Aubert and Droxler, 1996). The volcanic basement is cut by two deep grabens with a NNW-SSW orientation. In the Eocene, carbonate production was established on the topographic highs, and grabens served as seaways separating the shallow banks (Part I, this volume). Carbonate banks continued to accrete in the Eocene and early Oligocene. By the end of the early Oligocene time, the grabens were substantially filled with pelagic sediments and material eroded from the surrounding highs, and became dormant. The paleobathymetry data from an industrial well indicates that a significant sea-level fall at the early-late Oligocene transition (~28.5 Ma) exposed the tops of the carbonate banks
(Aubert and Droxler, 1996; Part I, this volume). The banks recovered in the late Oligocene and continued to aggrade and backstep, keeping pace with rising sea level. The seaways were partially filled with periplatform sediments.

In the late Oligocene, the banks began developing elevated rims along their margins that separated bank interiors from open waters. Numerous patch reefs were growing in the lagoons protected by the platform rim. In the early Miocene, the relative sea-level rise continued and bank interiors drowned. The bank rims, however, were able to keep up with the rapid sea level rise and continued to aggrade creating characteristic “empty bucket” geometry. Where possible, the banks backstepped onto higher ground and migrated upslope. Ultimately, a series of flat top banks were established at the periphery of the basin at the end of the early Miocene. During the middle Miocene, the bank margins progradated towards the central part of the basin. In the late Miocene and Pliocene, the central part of the basin was filled with periplatform sediments while the flat-top carbonate banks aggraded.

The present day atoll configuration is explained by frequent rapid sea level falls and rises that periodically exposed and flooded the flat-topped carbonate banks during the Quaternary. Periodic subaerial exposures resulted in carbonate dissolution by meteoric waters and the formation of karst topography. Carbonate production resumed during periods of flooding and the antecedent karst topography defines the shape of modern atolls and their lagoons (Purdy and Bertram, 1993).
DATA SET

Seismic data

The seismic data set used in this study comprises 6000 km of 2-D multi-channel seismic profiles acquired by Royal Dutch/Shell Oil Company. The seismic grid covers the entire Inner Sea area but does not extend beyond its limits into the atoll lagoons. The area covered by seismic forms a 275 by 50 km rectangle (Fig. II.3). The seismic data was collected in 1989 by GECO and processed by CGG in 1989-90. The average line spacing is 2 km between the east-west lines and 4 km between the north-south lines. The 60-fold multi-channel data was collected with four air gun arrays with an individual 4804 in$^3$ gun capacity, towed at a depth of 6 m and recorded using a 3000-m long 240-channel streamer with 12.5-m group spacing and 2-ms sampling rates. The shots were fired at a 25-m interval. The recording length of data was 6 seconds. Seismic processing included spherical divergence and geometrical spreading compensation, filtering, predictive deconvolution, velocity analysis at every 2 km, NMO correction, phase compensation, and time migration. The data quality of the zero-phase seismic lines is good to excellent down to 2 seconds, whereas it varies at deeper levels due to the ray-path problems under the carbonate reefal margins. A few seismic sections were sampled for frequency content and the measured dominant frequency at 1 second was 50 Hz, 35 Hz at 1.5 seconds and 25 Hz at 2 seconds. The vertical resolution for a zero-phase wavelet equals $\frac{1}{4}$ of the wavelength (Kallweit and Wood, 1982). Using the velocities from a VSP survey in the
Figure II.3. Shell 2-D multi-channel seismic (MCS) grid and locations of Elf NMA-1 and Shell ARI-1 wells and ODP Site 716. Thick lines indicate seismic lines discussed in text.
Shell ARI-1 well, the average vertical resolution would be 10 m at 1 second two-way travel time (TWT), 15 m at 1.5 seconds TWT, and 25 m at 2 seconds TWT.

Well data

Well data is crucial for the correlation and ground-truthing of seismic interpretations. The nature and age of sediments in the Maldives are known from two industrial wells, NMA-1 and ARI-1, and ODP Site 716.

NMA-1 well was drilled in 1976 by the Elf Aquitaine operated consortium in the North Male Atoll lagoon in 45 meters of water (Figs. II.2 and II.3). The well penetrated 2106 m of Eocene to Pleistocene carbonate sediments and bottomed in 116 m of weathered basalts (Fig. II.4). The well drilled through the bank margin and recovered late Oligocene – middle Miocene sediments. The well’s lithologic and stratigraphic record is incomplete due to significant gaps in sediment recovery caused by frequent losses of circulation while drilling and caving.

The ARI-1 well was drilled in 1991 by Shell in the central part of the Inner Sea in 348 m of water (Figs. II.2 and II.3). The well encountered 3315 m of late Eocene to modern carbonates and recovered 50 m of weathered basalts (Fig. II.5). Despite the fact that the upper 450 m of sediments (uppermost late Miocene and Plio-Pleistocene) were not recovered, and a few sample gaps exist in the lower part of the well record, the ARI-1 well provides reliable information on the composition, biostratigraphy, and paleobathymetry of the sediments in the basin. Biostratigraphic analyses of sidewall core
Figure II.4. Elf NMA-1 well drilled in the lagoon of North Male atoll and Shell seismic line E310NMA. Uneven sea floor bathymetry causes velocity pull-up under the atoll margin. See seismic line location on Figure II.3.
Figure II.5. Location of Shell ARI-1 well on the segment of seismic line E470 (see Figure II.3 for line location). The well was drilled to test the seismically defined dip-closed structure. The well's biostratigraphy and gamma ray log are shown. The seismic section was split to show location of the well.
and cutting samples included foraminifera, nannoplankton and palynomorphs. The well was logged using a standard suite of logging tools (caliper, gamma ray, resistivity, sonic, spontaneous potential, and density). A vertical seismic profile (VSP) provided seismic velocities essential for depth conversion and correlation between the seismic and well data.

ODP Site 716 (Figs. II.2 and II.3) was drilled in 544 m of water during Leg 115 in 1987 and recovered 264 m of periplatform carbonate sediments of the late Miocene to Recent age (Backman et al., 1988; Droxler et al., 1990; Malone et al., 1990). The sediments were represented by foraminifera-rich calcareous oozes with significant amount of bank-derived aragonite needles (Droxler et al., 1990; Malone et al., 1990). ODP Site 716 record complements the information from the industrial wells where the uppermost sections were not recovered.
MIDDLE MIocene PROGRADATION IN THE MALDIVES

Previous studies based on the Elf data

Aubert and Droxler (1992) described the middle Miocene progradation from both east and west towards the central part of the Inner Sea using seismic data in the northern part of the Inner Sea. They related the onset of the bidirectional progradation to the seasonal switch in the wind patterns caused by the establishment of the Indian monsoons. Purdy and Bertram (1993) further documented the regional bidirectional progradation in the Maldives. They reported different styles of progradation for the eastern and western rows of atolls. According to Purdy and Bertram (1993), the progradation under the eastern row of atolls is expressed by a progression from mounded to shingled to sigmoidal patterns. They stressed the greater extent of progradation from the east than from the west. The progradation style for the western row varied locally but typically was represented by shingled reflectors. Purdy and Bertram (1993) also reported the absence of progradation under the Felidu atoll in the eastern row.

Aubert (1994) and Aubert and Droxler (1996) described the bidirectional progradation in the Inner Sea as starting in the late middle Miocene and continuing until the late Pliocene. Maximum progradation occurred in the late middle Miocene-early late Miocene. The progradation was described as a series of flat-topped sigmoidal sequences attached to early Miocene "nucleating mounds". Aubert and Droxler (1996) also concluded that progradation was more pronounced for the western row rather than for the
eastern row of atolls. On some seismic profiles, individual sequences separated by downward shifts of onlaps where defined but "no coherent regional sea-level signal could be obtained" due to correlation difficulties (Aubert and Droxler, 1996, p. 518). On one seismic profile (Aubert and Droxler, 1996; their Figure 13), nine high-order sequences were defined. The sequences showed an evolution from low-angle aggrading ramps to an increasingly steeper prograding margins. The interpretation of stratigraphic geometries suggested a long-term regression punctuated by short-term transgressions and regressions (Aubert and Droxler, 1996).

**Interpretation of Shell data**

The interpretation of the more detailed seismic grid shows that a series of flat-top carbonate banks ("nucleating mounds" of Aubert and Droxler, 1996) were established in the peripheral parts of the basins at the end of the early Miocene (Part I, this volume). These banks owe their location to the dramatic backstepping of platform margins in the late Oligocene and the early Miocene (Fig. II.6) in response to a rapidly rising relative sea level. In the latest early Miocene, carbonate flat-top banks aggraded vertically by as much as 250 ms TWT (Fig. II.7). The deep seaways related to the basement-controlled topography were substantially filled with periplatform sediments in the late Oligocene and particularly in the early Miocene. By the end of the early Miocene, a relatively wide (~30 km) shallow central basin was formed as a result of the backstepping, partial drowning, and aggradation of the early Miocene banks.
Fig. II.6. Late Oligocene - early Miocene backstepping of the carbonate bank margins and the establishment of the late early Miocene flat-top banks. Interpreted segment of line E120.
The seismic lines imaging the Miocene carbonates have a few minor imaging problems. Channels in the shallower section create a velocity pull-down effect that propagates deeper into the section. This effect distorts the shape of some of the reflectors forming the middle Miocene clinoforms. The velocity effect was taken into account during the interpretation of stratial geometries.

Age control on sequence development

The timing of the progradation in the Maldives was established by correlation to the wells where the age of sediments was determined from micropaleontologic analyses of sidewall cores and cuttings. The Shell ARI-1 well biostratigraphy record is more complete than that from the Elf NMA-1 well, and correlation of seismic reflectors to ARI-1 is straightforward. The ARI-1 well, however, was drilled in the central part of the basin and recovered only the reduced, basinal counterpart of the prograding sequences (Fig. II.5). Because of this "condensed" stratigraphy in the well, only limited age control of the individual sequences is possible. Nevertheless, the timing of the initiation and development of progradation is fairly well constrained. The Shell/Worldwide foraminifer zones (Van Morkhoven and Schroeder, 1986; Styzen, 1996) were used to define the biostratigraphy of the section. These zones are based on the last appearance datums (LADs) instead of first appearance (Blow, 1979). LADs are defined as the first encounter of the species downwhole, which has a practical meaning when working with well cuttings. The base of the prograding sequences was dated as early middle Miocene
(Shell/Worldwide zone SN10-12). The top of the last prograding sequence attached to a slope of the early Miocene flat-top bank was defined between the zones SN14 and the base of zone SN15. This establishes the timing of the last sequence formation as latest middle Miocene or the earliest late Miocene. Progradation of bank margins continued locally until the end of the early Pliocene.

Prograding complexes

The progradation of the bank margin in the Maldives began in the middle Miocene. The progradation occurred from both east and west towards the central part of the basin. The areas of progradation are easily identified on the isopach map of the middle Miocene unit (Fig. II.8) as areas of local thickness increase. Thick bulges represent depocenters where prograding clinoforms were stacked. The name “prograding complexes” was chosen to describe such areas of “laterally extensive and vertically significant set of two or more clinoforms”. This term should not be confused with the “prograding complex” of Mitchum et al. (1994) who restricted this term to describe a lowstand prograding wedge.

Three prograding complexes were identified within the limits of the Shell seismic grid (Fig. II.8). Complex I is located in the northwestern corner of the grid near Horseburg atoll. In the map view, the depocenter has an oval shape with dimensions of 20 by 30 km. Fifteen Shell seismic lines offer different cross sectional views through the clinoforms of Complex I. Prograding Complex II is located in the west-central part of the
Middle Miocene progradation based upon the interpretation of Elf seismic data (Aubert, 1994; Aubert and Droxler, 1996)

Figure II.8. Isopach map of the middle Miocene prograding sequences.
Inner Sea between Ari and North Nilandu atolls. The complex covers an area of 20 by 30 km and it is imaged on twenty Shell seismic lines. Prograding Complex III is located in the northeastern part of the Shell seismic grid north of Gaha atoll and is imaged on only three Shell seismic lines. Progradation was also reported, outside the area covered by Shell seismic lines, on the Elf lines acquired in the atoll lagoons and in the northern part of the Inner Sea (Aubert and Droxler, 1992, 1996; Purdy and Bertram, 1993; Aubert, 1994).

The shaded areas on Figure II.8 indicate locations where the middle Miocene progradation was previously documented by Purdy and Bertram (1993) and Aubert and Droxler (1996). As shown by Aubert and Droxler (1996), the progradation from the western side appears to have been more substantial than from the eastern side, in both the thickness of sediments and the lateral outbuilding of the margins.

Seismic expression of prograding sequences

Five depositional sequences are defined within each prograding complex on the basis of the stratigraphic relationships, variation in seismic facies, and stratal geometries of the seismic units. The sequences have clinoform shape in dip view and thin out in the basinward direction to be commonly represented by a single reflector (Figs. II.9A and II.9A). The sequence boundaries were named A through E starting from the older sequence and are identified on seismic lines (Fig. II.9B).
Fig. II.9A. Uninterpreted segment of Line E130 showing middle Miocene clinoforms.
Figure II.9B. Interpreted segment of Line E130 showing five middle Miocene prograding sequences (numbered 1-5). Letters indicate sequence boundaries. Each sequence is divided into strong amplitude reflector packages (SARPs shaded areas) and weak amplitude reflector packages (WARPs, no color fill). Also shown are late Miocene sequences with onlapping wedges forming the lower part of the sequences.
Each of the five middle Miocene sequences can be divided into two seismic packages: a strong amplitude reflector package (SARP) that forms the lower part of the sequence, and a weak amplitude reflector package (WARP) which comprises the upper part of the sequence. SARPs consistently display basinward shift in onlap relatively to the underlying seismic package. The internal reflectors comprising SARPs are typically oblique or sigmoidal-oblique. The lower SARP reflectors onlap on the sequence boundary and the shallower reflectors that form the topsets of the SARPs display either toplap or downstepping (onlap onto the previous reflector at a lower level). The reflectors forming clinoform toesets downlap onto the sequence boundary. On seismic profiles that cut through the strata at an angle to the depositional dip, SARPs have a convex-upward shape (Figs. II.9A and II.9B). The topsets of the weak amplitude reflector packages (WARPs) that form the upper part of the prograding sequences extend in a shoreward direction and progressively onlap on the sequence boundary. The topsets of the uppermost WARP reflectors display toplap and the bottomsets have downlap terminations.
ANATOMY OF PROGRADING COMPLEX I

Prograding Complex I, located in the NW corner of the Inner Sea, is imaged on the dense grid of 2-D seismic lines showing its internal architecture in detail. The prograding clinoforms are attached to the edge of an early Miocene bank that is clearly displayed on line E120 (Fig. II.10). Unfortunately, line E120, as well as other Shell lines, do not extend far west enough to image the bank in its entirety. This bank, however, is imaged on the Elf seismic profiles (Aubert and Droxler, 1996, their Figure 13). The carbonate bank is characterized by a flat top and a relatively gentle slope front (~7 degrees). The bank slope is represented by basinward-dipping reflectors. Horizon A was picked as the base of the prograding sequences (Fig. II.9B) and was correlated around the basin. The time-structure map of Horizon A (Fig. II.11) partially shows the location of the bank edges at the periphery of the Shell seismic grid and the central trough at the beginning of the middle Miocene.

A detailed description of stratal geometries of prograding Complex I is possible from the seismic grid. Table II.1 summarizes the main parameters of prograding sequences of Complex I. The parameters include maximum sequence thickness, area and orientation of the depocenter, and maximum angle of the clinoform front. An average sonic velocity of 2500 m/s was used in order to convert the seismic data to depth. This value was determined by analysis of stacking velocities and the comparison of seismic facies with facies of known velocities from the VSP survey in the ARI-1 well.
Figure II.10. Segment of Line E120 showing middle Miocene prograding sequences (1-5) and the edge of the early Miocene flat-top carbonate bank. Letters identify sequence boundaries.
Figure II.11. Time-structure map of sequence boundary A (top of the early Miocene). The map shows the central trough and margins of flat-top banks on its periphery.
Table II.1. Parameters of the middle Miocene prograding sequences.

<table>
<thead>
<tr>
<th>Sequence</th>
<th>max thickness, ms TWT</th>
<th>max thickness, m</th>
<th>Area, km²</th>
<th>max angle, degrees</th>
<th>orientation of the depocenter</th>
<th>lateral migration of the depocenter, km</th>
</tr>
</thead>
<tbody>
<tr>
<td>SARP1</td>
<td>220</td>
<td>275</td>
<td>110</td>
<td>7</td>
<td>N-S</td>
<td></td>
</tr>
<tr>
<td>WARP 1</td>
<td>128</td>
<td>160</td>
<td>86</td>
<td>4</td>
<td>N-S</td>
<td>3</td>
</tr>
<tr>
<td>2</td>
<td>125</td>
<td>156</td>
<td>107</td>
<td>6</td>
<td>NNE-SSW</td>
<td>1.5</td>
</tr>
<tr>
<td>3</td>
<td>107</td>
<td>134</td>
<td>128</td>
<td>6</td>
<td>NNE-SSW</td>
<td>2</td>
</tr>
<tr>
<td>4</td>
<td>154</td>
<td>192.5</td>
<td>93</td>
<td>7.5</td>
<td>NNE-SSW</td>
<td>2</td>
</tr>
<tr>
<td>5</td>
<td>166</td>
<td>208</td>
<td>126</td>
<td>9</td>
<td>NE-SW</td>
<td>3</td>
</tr>
</tbody>
</table>

Table II.1. Main parameters of the middle Miocene prograding sequences. TWT values were converted to depth using 2,500 m/s sonic velocity.
Sequence 1

Horizon A forms the base of the prograding Sequence 1 (Fig. II.9B). The lower part of the Sequence 1 SARP is characterized by undulating and mounded reflectors on some of the dip lines (Figs. II.9 and II.10). A U-shape seismic unconformity, displayed on the N-S seismic line N040 (Fig. II.12), is interpreted as a slide scar created by bank slope collapse. The slide created a 6-km-wide breach that later filled with sediments of Sequence 1. The sediments above Horizon A appear as chevron-shaped seismic facies interpreted as rotated blocks (Fig. II.12). The detachment surface is listric and concave-upward in the cross-sectional view (Figs. II.9 and II.10), and likely has an amphitheater-like shape in map view. Gravitational collapse structures of similar geometry have been reported from different settings and occur at various scales (e.g. Hesthammer and Fossen, 1999). The causes of such gravitation collapses include seismic shocks, oversteepening of slopes, rapid rates of sedimentation and changes in pore fluid pressure.

The area of the collapse feature is coincident with limited area of thickening on the Sequence 1 SARP isopach map (Fig. II.13). This example illustrates creation of new accommodation space by the local removal of sediment. Sequence 1, however, is not limited to the filling of the space created by the slide and has a more regional character. Figure II.13 shows that SARP 1 is up to 275 m thick locally (Table II.1) in the area where the sediments filled the space created by the gravitational sliding. The sediments of Sequence 1 WARP form a wedge with a maximum thickness of 160 m (Fig. II.14). The comparison of the SARP 1 and WARP 1 maps shows that the locus of deposition shifted 3 km to the east from its previous position (Fig. II.15).
Figure II.12. Segment of Line N040 showing the strike view of the gravitational slide.
Figure II.14. Isopach map of Sequence I WARP.
Figure II.15. Depocenters of the middle Miocene prograding sequences of Complex 1.
Sequence 2

Horizon B forms the sequence boundary separating Sequences 1 and 2 (Fig. II.9B). On line E130 (Figs. II.9A and II.9B), Sequence 2 SARP appears as a convex-upward body with oblique internal reflectors which is shifted significantly (~4 km) in the basinward direction relatively to the topset of WARP 1. The vertical component associated with this downward shift is 87.5 m (Table II.2). The magnitude of this shift, however, appears to be smaller (77.5 m) on line E120 (Fig. II.10). The magnitude of the downward shift on line E130 is overestimated as a result of the line cutting at an angle oblique to the true dip. The overlying reflectors of Sequence 2 WARP onlap onto the Sequence 2 SARP and the sequence boundary B. The maximum angle of the clinoform foresets observed in Sequence 2 is 6 degrees, and the maximum thickness of Sequence 2 is 156 m (Table II.1). The axis through the thickest area (depositional center of sedimentation) of Sequence 2 shifted 1.5 km in the southeast direction relatively to the depositional center of Sequence 1 LARP (Fig. II.15). The orientation of the prograding wedge depositional center in map view changed from N-S to a NNE-SSW orientation.

Sequence 3

Horizon C forms the base of Sequence 3 and was dated by correlation to the AR-I-1 well (Fig. II.9B), where a sample from just above Horizon C was placed in the Shell/worldwide foraminifer zone SN13 indicating a late Serravallian age (11.75-11.5 Ma). Sequence 3 SARP is thicker and laterally more extensive than the SARP of Sequence 2. The topset of SARP 3 is shifted about 2 km basinward relatively to the topset of WARP 2 and the vertical component accounts for 62.5 m on line E130 and 50 m
Table II.2. Anchor points values measured on three seismic lines (E120, E130, and E140) crossing prograding Complex I, and the stacked relative sea-level record. Shaded cells mark the lowest TWT values that represent the shallowest position of the strata. These values were used to produce the stacked record.

| Line E120 | | Line E130 |
|---|---|---|---|---|---|---|---|
| anchor point | depth TWT, ms | magnitude TWT, ms | magnitude, m | anchor point | depth TWT, ms | magnitude TWT, ms | magnitude, m |
| bank top | 764 | | | bank top | 760 | | |
| SARP 1 onlap | 792 | -24 | -62 | 814 | -54 | -67.5 |
| | 768 | 24 | 30 | 802 | 12 | 15 |
| SARP 2 onlap | 830 | -62 | -77.5 | 872 | -70 | -87.5 |
| | 774 | 56 | 70 | 788 | 84 | 105 |
| SARP 3 onlap | 814 | -40 | -50 | 838 | -50 | -62.5 |
| | 770 | 44 | 55 | 768 | 70 | 87.5 |
| shelf break | 858 | -88 | -110 | 826 | -58 | -72.5 |
| | 846 | -88 | -110 | 846 | -20 | -25 |
| | 866 | -88 | -110 | 866 | -20 | -25 |
| coastal? onlap | 764 | 94 | 117.5 | 788 | 78 | 97.5 |
| | 852 | -88 | -110 | 874 | -86 | -107.5 |
| | 760 | 92 | 115 | 770 | 104 | 130 |
| aggradation | 674 | 86 | 107.5 | 720 | 50 | 62.5 |
| aggradation | 606 | 66 | 82.5 | 698 | 22 | 27.5 |

| Line E140 | | Stacked magnitude, TWT ms | | Stacked magnitude, TWT ms |
|---|---|---|---|
| anchor point | depth TWT, ms | magnitude TWT, ms | magnitude, m | Stacked magnitude, TWT ms | Stacked magnitude, m |
| bank top | 796 | | | 0 | 0 |
| | | | | -32 | -40 |
| | | | | 24 | 30 |
| | | | | -62 | -77.5 |
| | | | | 56 | 70 |
| SARP 3 onlap | 869 | -24 | -30 | 40 | -50 |
| | 810 | 59 | 73.75 | -46 | 57.5 |
| SARP 4 onlap | 820 | -10 | -12.5 | -52 | -65 |
| | | | | -26 | -32.5 |
| SARP 4 onlap | 764 | -64 | -80 | -20 | -25 |
| | 790 | 30 | 37.5 | 102 | 127.5 |
| SARP 5 onlap | 840 | -50 | -62.5 | -76 | -95 |
| | 792 | 48 | 60 | 80 | 100 |
| aggradation | 770 | 22 | 27.5 | | |
| aggradation | 676 | 94 | 117.5 | | |
on line E120 (Figs. II.9B and II.10). The overlying LARP 3 is 70 ms TWT (87.5 m) in its thickest part. The isopach map of Sequence 3 shows further migration of the progradation depocenter towards a southeast direction (Fig. II.16).

**Sequence 4**

Horizon D forms the boundary between Sequences 3 and 4. The internal reflectors of Sequence 4 SARP on line E130 show systematic downstepping of the bank margin (Fig. II.17). A series of erosional unconformities are arranged in a step-like fashion at progressively lower topographic levels. This geometry is commonly referred to as "erosive regression" (Curry, 1964) or "forced regression" (Posamentier et al., 1990; Posamentier et al., 1992; Hunt and Tucker, 1992, 1995). Sequence 4 imaged on Line E120 (Fig. II.10) does not display the downstepping of the shelf margins, possibly due to the oblique orientation of the section. Within the sequence, however, a reflector with a steep break is interpreted as an erosional cliff (Fig. II.18). On Line 140 (Fig. II.19), the lower part of Sequence 4 is expressed as series of downstepping SARPs, and possibly a slumped SARP. The reflectors of Sequence 4 WARP onlap onto the previous sequence. The uppermost oblique reflectors form a small prograding wedge (Fig. II.9B). The maximum thickness of Sequence 4 (Fig. II.20) is estimated as 192.5 m (Table II.1). The slope of the bank margin in Sequence 4 is steeper than in previous sequences and reaches 7.5 degrees. Sequence 4 is also characterized by a well-defined slope break that contrasts with the smoother slopes in preceding sequences. The height of the slope, or vertical distance between the clinoform break-point and the basin floor, is 320 ms TWT (400 m).
Figure II.16. Isopach map of the middle Miocene Sequence 3.
Figure II.17. Uninterpreted (A) and interpreted (B) segment of Line E130 showing downstepping margins ("forced regression") geometry of the middle Miocene Sequence 4.
Figure II.18. Uninterpreted (A) and interpreted (B) segments of Line E120 showing internal geometries of middle Miocene sequences 1-5.
Figure II.19. Segment of Line E140 showing the middle Miocene progradling sequences (1-5). Shaded areas represent SARPs and areas with no color fill represent WARPs.
Figure II.20. Isopach map of the middle Miocene Sequence 4.
Sequence 5

Horizon E marks the sequence boundary between Sequences 4 and 5 (Fig. II.9B). The stratal geometries in this sequence are unique. First, convex-upward reflectors with bi-directional downlap onto the Horizon E are observed at the distal part of the clinoform slope toe (Figs. II.9B and II.19). The reflectors represent mound-like bodies which bear striking resemblance to basin floor fans, as described in the siliciclastic deepwater settings (e.g. Van Wagoner et al., 1990; Mitchum et al., 1994). Individual mounded features are up to 50 m (using 2500 m/s sonic velocity) and 2 km wide (Figs. II.9B and II.19). Their base typically rests on the flat surface of horizon E. On line E130 (Fig. II.9B), the mounds are located 6 km west of the Sequence 4 clinoform break. Some individual mound-like features appear to be joined together or overlap with one another, others are separated by V-shaped features that potentially represent channels. It is important to mention that the basin-floor mounded features are unique to Sequence 5 and are present across the whole basin.

The mounded features of Sequence 5 are covered by an apron of sediments onlapping onto the sequence boundary E (Fig. II.9B). SARP\s are absent on lines E120 and E130 but a well-defined SARP is present on line E140 (Fig. II.19). On line E130 (Fig. II.9B), a well-developed shelf-slope break is recognized. The upper part of sequence 5 shows oblique reflectors that form a prograding wedge. The isopach map for Sequence 5 (Fig. II.21) clearly shows that thicker sediments were deposited at the toe of the bank slope. The maximum thickness of Sequence 5 is 208 m (Table II.1).
Figure II.21. Isopach map of the middle Miocene prograding Sequence 5.
**Strike view**

Prograding clinoforms are easily recognized in dip oriented seismic and outcrop sections but their appearance in depositional strike view has not been well documented. Shell seismic line N060 provides a view close to strike for complex I (Fig. II.22). In this cross section, the clinoforms appear as broad mounds with numerous small channels oriented perpendicular to the margin. The channels are present both at the sequence boundaries and within the sequences at different levels (Fig. II.22). The width of the channels ranges from 150 to 700 m (with an average width of 300 m), and the depth of the channels varies from to 25 to 75 m. Many channels display multiple cut-and-fill events. These channels likely served as the conduits that brought the carbonate material from the bank top to the outbuilding clinoform fronts. This configuration demonstrates that during progradation the sediments were delivered by a line source rather than supplied by a single point source. The strike view also shows the migration of the sedimentation depocenter through time. On line N060 (Fig. II.22), the depocenter migrated from north to south from the time of Sequence 1 deposition for approximately 20 km.

**Overlying sequences**

Vigorous progradation is halted at the end of the middle Miocene and is followed by a transgressive filling of the basin. The overlying upper Miocene sequences 6 and 7 (Fig. II.9B) display local progradation of bank margins and filling of the central basin. The reflectors of the lower part of Sequence 6 form a wedge onlapping onto the sequence boundary F. The overlying reflectors continue progressively onlapping onto the toesets of
Figure II.22. Segment of Line N060 showing strike view of the middle Miocene prograding Complex I.
sequence 5. The onlap is likely marine and not coastal. Sequences 6 covers entire Complex I. Reflectors of Sequence 7 onlap onto the sequence boundary G and represent the continued filling of the central trough.
INTERNAL ARCHITECTURE OF PROGRADING COMPLEXES II AND III

Prograding Complex II

Prograding Complex II, which occupies the area between Ari and North Nilandu atolls, is imaged on twenty Shell seismic lines (Fig. II.8). Unfortunately, Shell seismic lines do not extend west far enough to image the edge of the flat-top bank to which the complex is attached (Fig. II.23). This prevents relative sea-level reconstruction based on the interpretation of stratal geometries of Complex II because it does not allow direct measurement of basinward shifts and coastal onlaps.

Five sequences were identified in Complex II. The reflectors that mark the base of Sequences 1 and 2, and the top of Sequence 3 in Complex I, were correlated across the basin and form corresponding sequence boundaries in Complex II. The lower three sequences may be divided into SARPds and WARPds (Fig. II.23), just like the sequences in Complex I. Sequence 4 is significantly thinner than Sequence 4 from prograding Complex I, where force-regressive geometry was observed on seismic line E130 (Fig. II.17). Mound-like features, interpreted as gravity-flow deposits, are found at the base of Sequence 5 in the prograding Complex II. Deposition of these features is synchronous with the deposition of the gravity-flow deposits found in Sequence 5 in the prograding Complex I.
Figure II.23. Interpreted segment of line E640 through the middle Miocene prograding Complex II. Shaded areas represent SARPs, areas with no color fill represent WARPs.
The number, internal architecture, and seismic character of prograding sequences of Complex II therefore matches, with minor differences, the sequences of Complex I. Seismic correlation suggests that the deposition of prograding sequences in Complex II was synchronous with the deposition of sequences in Complex I. Margins of the banks separated approximately 150 km from each other, simultaneously produced similar stratal geometries. This implies that deposition of the middle Miocene prograding sequences was driven by a mechanism of regional scale.

**Prograding Complex III**

Prograding Complex III north of Gaha atoll and is imaged on only three Shell seismic lines. Line E130 imaged the edge of the flat-top bank to which Complex III is attached (Fig. II.7).

Prograding Complex III is located approximately 30 km to the east from prograding Complex I. Five prograding sequences were identified within Complex III (Fig. II.7). Sequence 1 is characterized by a basinward shift in onlap. The lower part of Sequence 2 has forced-regressive geometry, with four downstepping wedges (Fig.II.7). Sequences 3, 4, and 5 in Complex III are thinner than in Complex I, which can be explained by an oblique imaging of the sequences in Complex III. Differentiation into SARPs and WARPs in the sequences of Complex III is not obvious, which again may be attributed to the limited coverage by seismic profiles and oblique imaging.
Correlation of seismic reflectors marking sequence boundaries in prograding complexes I and III is straightforward and shows that deposition of sequences in different complexes was synchronous (Fig. II.24). This further justifies that progradation of bank margins in the Maldives in the middle Miocene was simultaneous, and therefore driven by a regional-scale process.
Figure II.24. Interpreted Shell seismic line E120. Compaction-driven differential subsidence causes distortion of the middle Miocene sequences.
INTERPRETATION OF SEISMIC FACIES AND STRATAL
GEOMETRIES

SARPs and WARPs

The difference in the amplitude of internal reflectors between SARPs and WARPs within prograding sequences may be explained by differences in lithology and/or diagenetic overprint. The actual composition of the prograding sequences is unknown because wells did not penetrate the prograding clinoforms. Seismic properties of carbonate sediments and rocks depend on a wide range of parameters including porosity, pore type and shape, pore fluid, and saturation (Wang, 1997), with pore type being most important. The reflectivity of sediments is caused by changes in acoustic impedance, which is a product of sediment sonic velocity and density. Anselmetti (1994) showed that good reflectivity of carbonate sediments in seismic sections was commonly caused by variations in fabric of carbonate sediments and not by the intercalations of carbonate and non-carbonate sediments, as often thought.

Diagenetic alteration is one of the most important factors influencing the sonic velocity of carbonates. Coarse-grain sediments are more susceptible to diagenetic alteration and reach higher sonic velocities after shallow burial faster than fine-grained sediments (Anselmetti, 1994). SARPs forming the lower part of the prograding sequences are likely to be composed of beds with relatively coarse grains, different grain-to-mud
ratio, and a degree of diagenetic alteration. WARP\textsubscript{s} are likely to be composed of more homogeneous sediments and are probably more mud-rich, based on their overall poor reflectivity.

Based on the interpretation of stratal patterns, stratigraphic relationships, and seismic facies, SARP\textsubscript{s} are interpreted as sediments deposited during periods of falling relative sea level. Based on the assumption that all carbonate sediment is produced locally on the flooded bank top at sea level, it is assumed that a basinward shift in onlap represented by SARP\textsubscript{s} is equivalent to the downward shift in coastal onlap. Falling base level resulted in the exposure of the preceding sequence topsets and downward (basinward) migration of the bank margin. This downstepping, or forced regression, is clearly expressed on some of the seismic profiles (e.g. Figs. II.7 and II.17). This characteristic geometry is a solid indicator of falling relative sea level because neither variation in sediment supply nor environmental changes may be responsible for the downward migration of the bank margin (Schlager, 1992). The sediments that form SARP\textsubscript{s} were probably winnowed by currents and have a high grain-to-mud ratio due to the removal of fine-grained material. Because of their high primary porosity, these sediments are also likely to be well and more quickly cemented and consequently have high sonic velocity values than the overlying WARP\textsubscript{s}.

WARP\textsubscript{s} were interpreted as sediments deposited during relative sea rise level and subsequent highstand. The onlapping strata of the lower part of WARP\textsubscript{s} represent the initial flooding after the preceding base level drop. Flooding results in a greater area available for sediment production, and an increase in the amount of muddier sediment
produced by the carbonate factory. The prograding uppermost part of a WARP represents deposition during the sea level highstand when excess sediment is shed from the bank top.

**Interpretation of the Middle Miocene Prograding Sequences**

Each prograding sequence consists of two packages: a basal, SARP, and an upper, WARP. Each sequence therefore represents a complete sea level cycle. The cycle begins with a relative sea level fall during which an unconformity is developed and carbonate production is reduced and shifted in a basinward direction. The subsequent flooding causes the backstepping of the bank margin and an increase in the amount of carbonate sediment produced. During the last part of the cycle, a relative sea level highstand, the bank margin progrades and the bank top sheds the excess material.

In the middle Miocene prograding complexes of the Maldives, the vertical component of downward shifts represented by SARPs appears to be more significant than the vertical aggrading component of WARPs. This observation indicates that the sea level falls had greater magnitude than the intermittent rises. The early Miocene flat-top banks did not aggrade when the bank edge was prograding, which implies that the middle Miocene relative sea level never rose back to the position of the early Miocene.

Sequence 5, the terminal middle Miocene prograding sequence, deserves special attention. The slope of the prograding margin has become progressively steeper and reached its maximum declivity at the time of Sequence 5 formation. Steepening of the
prograding margins through time has been reported from different localities around the world (e.g. Eberli and Ginsburg, 1989; Pomar, 1993; Tinker, 1998). In the case of the Maldives middle Miocene prograding margins, the steepening of the clinoform slope was possibly related to a significant sea-level fall during the deposition of Sequence 4, and resulting forced-regressive geometry. Erosion due to exposure and the formation of narrow shelves during the falling base level led to the creation of a steeper margin. The slope of the overlying sequence 5 was developed on the antecedent steep slope and continued to steepen until it failed.

The deepwater mounded lobe-shaped features at the bottomsets of the clinoforms in Sequence 5 are interpreted as gravity-flow deposits. Their deposition is connected to the steepening of platform margins and their failure. The channels on top of the lobes likely played the distributary function and served as the conduits of the carbonate material. It does not seem possible, however, to conclude whether the channels originated on the margin slopes or not. The mapping of individual lobes and channels is not possible at this time because their size is typically smaller than half-spacing of seismic lines. It is also difficult to determine whether the formation of the basin mounded features occurred as a result of a single or multiple events. It is important to note that the lobes of Sequence 5 are found throughout of the paleo-Inner Sea basin floor, but only within Sequence 5. This implies a regional and unique set of conditions for their formation and possibly a common trigger mechanism that facilitated their deposition.

The deep-water mounded deposits of Sequence 5 may be similar to the late Cretaceous coarse-grained lobes described by Eberli et al. (1993) in the Maiella Platform
in Italy. The lobes described in Maiella had a positive relief of up to 70 m and extended laterally for up to 2 km. Internally, the lobes consisted of stacked channel complexes. These coarse-grained lobes were interpreted as lowstand slope fan deposits (Eberli et al., 1993).

In the Maldives, the mound-shaped seismic facies on the basin floor above sequence boundary E are also similar to the description of the Pleistocene turbidites and debris flow deposits in the Tongue of the Ocean (Schlager and Chemark, 1979) and in the Exuma Sound (Crevello and Schlager, 1980) in the Bahamas. These deposits formed coalescing mounded lobes at the toe-of-slope and basin floor. The turbidites in the Exuma Sound were mainly composed of the clean graded skeletal sands and lithoclasts, and the gravity flow deposits were comprised of pebbly mud with rubble and graded carbonate sands (Crevello and Schlager, 1980). Dating showed that these debris-flow deposits formed 75,000 to 80,000 years BP during the falling of sea level (Droxler, 1984).

Betzler et al. (1999) also described the mounded morphology of the Miocene turbidite lobes on the western side of the Great Bahama Bank. The turbidites lobes had flat to convex-down reflectors in their central part interpreted as turbidite feeder channels.

In siliciclastic systems, the deposition of gravity-flow deposits on the basin floor is associated with relative sea level falls (e.g. Van Wagoner et al., 1988). In the carbonate systems, however, the deposition of turbidites and debris flow deposits in the basin is not considered to be diagnostic to establish the relative sea level position. It has been demonstrated that the export of shallow water carbonate material in the form of turbidites
is more common during sea level high stands (Droxler and Schlager, 1985; Schlager et al., 1994). Under the “highstand shedding” scenario, the carbonate factory produces and exports a greater amount of material when the bank top is flooded. Late Quaternary high-stand turbidites typically are composed of aragonite and high-Mg calcite non-skeletal bank-derived material (Haak and Schlager, 1989). During sea level lowstands, a large part of the bank is exposed which results in a complete or partial shutoff of carbonate production. This exposure commonly results in sediment erosion and the redistribution of sediments downslope, in a form of coarse-grained debris flows, slumps, and megabreccias (Grammer et al., 1993; Handford and Loucks, 1993).

A study by Spence and Tucker (1997) showed that the probability of carbonate megabreccia occurrence is higher during sea level falls due to overpressuring in the confined horizons, as the pore-fluid drains from sediments created when the platform top becomes exposed. The link between the relative sea level fall and the development of overpressure becomes important when we consider the basinal counterparts of Sequences 4 and 5 in the Maldives. In the central and deepest part of the basin, Sequences 4 and 5 are represented by a seismic unit showing a lateral change in character from parallel internal reflectors to disrupted reflectors resembling imbrication (Fig. II.25). Aubert and Droxler (1996) first noted the seismic character of this unit. This basin equivalent of prograding Sequences 4 and 5 is up to 137 m (110 ms TWT) thick. Two possible explanations for this layer-bounded disturbance are considered. The first explanation relates the deformation to compression-related faulting which is gravity-driven. The second interpretation may be soft-sediment deformation related to early compaction of
Figure II.25. Interpreted segment of line E470 showing layer-bounded disrupted deposits of the middle Miocene sequences 4 and 5. Parallel continuous reflectors on the east grade into disrupted, "imbricated" reflectors to the west.
fine-grained sediments due to dewatering. Regionally extensive polygonal fault systems related to volumetric contraction during early dewatering have been documented in basins worldwide (Cartwright, 1996; Cartwright and Dewhurst, 1998; Dewhurst et al., 1999). This type of deformation is found in marine deposits composed of ultrafine-grained claystones or carbonate chalks.

The second explanation is favored because the gravity-driven compressional faulting should not be unique to one horizon and would be expected in strata both above and below the described unit. The mounded gravity flow deposits of Sequence 5 are thus interpreted as lowstand deposits related to a significant relative sea level fall. The sea-level fall also caused the overpressuring of the low-permeability fine-grained basin sediments that resulted in the geologically instantaneous fluid expulsion.
RECONSTRUCTION OF THE MIDDLE MIOCENE RELATIVE SEA LEVEL
HISTORY FROM STRATAL GEOMETRIES

Carbonate platforms as recorders of sea level change

Carbonate platforms are highly sensitive to relative sea level fluctuations because optimum carbonate production is confined to the euphotic zone (Bosscher and Schlager, 1992). Relative sea-level changes are result of interplay between accommodation space and sediment supply. Carbonate platforms respond to these changes by creating characteristic depositional geometries that can be recognized on seismic sections, in outcrop, and on well-log cross sections (Schlager, 1992, 1993). Accommodation space, or space available for sediment deposition, is created by interplay between tectonic subsidence and eustatic sea level fluctuations. The common assumption is that under the stable tectonic conditions (e.g. on passive margins), the rate of tectonic subsidence is an order of magnitude lower than the rates of the eustatic sea level falls and rises, and sea level fluctuations therefore play the major role in shaping sedimentary geometries (Vail et al., 1991).
Significance of the Maldives setting

The tectonic history and physiographic features of the Maldivian platform make it an ideal natural laboratory to study the carbonate system response to the sea-level changes. The position of the Maldives on the rigid Indian Plate and the paucity of faults in sediments younger than early Oligocene on the seismic sections demonstrate the near absence of synsedimentary tectonic activity in the middle Miocene (Part I, this volume). The large size of this isolated platform and the relatively shallow central internal basin facilitated the formation of stratigraphic geometries typical to those on passive margins. The remoteness from land and surrounding deep waters protected the Maldives from large pulses of terrigenous material. All carbonate production in the Maldives was local and controlled mainly by environmental factors. Such environmental factors include biota, water temperature and salinity, and nutrient supply (Schlager, 1991). The variations of such environmental factors within a platform are difficult to recognize in the geological record, in particular in an area such as the Maldives where the sampling of the sediments is so limited.

Methodology of relative sea level reconstruction

Traditionally, reconstruction of relative sea level from seismic data was based upon the interpretation of 2-D cross sections and consisted of measuring the magnitudes
of the shifts in coastal onlap (Vail et al., 1977). While the shifts themselves are fairly
easy to document, the magnitudes of these events are difficult to capture. By definition,
Type 1 sequence boundaries are formed when relative sea level falls below the offlap
break exposing the shelf (Vail et al. 1977). Downward shifts of coastal onlap record the
increase of the rate of sea level fall. Coastal aggradation indicates a relative sea level rise
that can be measured where littoral deposits onlap the underlying sequence boundary. A
number of problems, however, make the practical application of this technique difficult.
Coastal onlap, in many cases, is impossible to distinguish from marine onlap on seismic
sections if it is not comprehensively sampled by drilling. In many settings, sediments
onlapping onto the slope are a product of strong deep water drift currents (e.g. Marion
Plateau off northeastern Australia; western slope of the Bahamas platform, Anselmetti et
al., in press). Onlapping deepwater drift sediments interfinger with slope-margin system
and further complicate the sea level reconstruction.

The classic sequence stratigraphic model defines the offlap break as an indicator
of relative sea-level position (Vail et al., 1977). However, the actual physical meaning of
the offlap break and its paleowater depth is strongly debated (Steckler at al., 1999). The
original assumption (Vail et al., 1977) was that the offlap break occurs in 10-20 m of
water depth, at the fair weather wave base (FWWB). Recent attempts to test this
assumption in siliciclastic systems demonstrated that the paleowater depth at the
clinoform break might be not depth-characteristic. Fulthorpe and Austin (1998) estimated
the paleowater depth of the offlap break in New Jersey Margin Miocene section as 80 to
100 m, which greatly exceeds the FWWB. Backstrip modeling of the New Jersey margin
geometries led Steckler et al. (1999) to conclude that the clinoform break points were at a water depth of 100 meters, and that exposure of the entire shelf and shelf edge was not required to produce a type 1 sequence boundary. Steckler et al. (1999) concluded that sea-level estimates based on the assumption that onlap over the clinoform front was coastal were unreliable.

Eberli and Ginsburg (1989) measured the vertical aggradation and downward shifts between the clinoform break points in the Bahama platform to estimate the episodic sea level fluctuations in the Miocene-Pleistocene. While this approach is valuable, a large error may be introduced. First, there is uncertainty about the paleowater depth of the clinoform breakpoint, which may be different for different clinoforms. Second, the position of the clinoform breakpoint is easily defined on the steep slopes but is more difficult to determine on the low-angle slopes with small curvature.

Sonnenfeld and Cross (1993) used the aggradation:progradation ratios and offlap angles to quantify volumetric partitioning in the Permian carbonate strata of the San Andres Formation of Last Chance Canyon, New Mexico. Tinker (1998) quantified the stratigraphic evolution of the Upper Permian Capitan depositional system (New Mexico and Texas) by calculating depositional parameters such as: progradation and aggradation (and associated offlap angle) of the shelf-crest and shelf-margin facies tract; distance from the shelf crest to reef; reef depth; outer-shelf dip angle; and lateral distance and depth from the shelf crest to the toe of slope. The application of this approach to the middle Miocene strata in the Maldives is difficult. First, the middle Miocene progradation in the Maldives did not occur on a steep rimmed margin and the slope-shelf break is not
well defined in many of the prograding sequences. Second, there is no evidence for the presence of a reef complex within the sequences.

An important problem of sea level reconstruction from stratal geometries is unfilled accommodation space. Eberli (1999) pointed out that in the modern and recent carbonate environments the available accommodation space is not completely filled with sediments. This implies that the amount of sediment aggradation will give an underestimated measurement of the increase of the accommodation space. There are certain platform elements, however, such as flat top banks, shoals, and reef margins that closely follow the sea level position and are diagnostic for the reconstruction of the past sea level position.

Anchor points

In this study, I attempt to reconstruct the relative sea level or accommodation history from the stratal geometries of the middle Miocene prograding complex and estimate the timing and amplitudes of these events. A method to measure the amplitudes of relative sea level in a carbonate system with little or no siliciclastic input is proposed here. The reconstruction of the sea-level position is based on a set of "anchor points", or diagnostic geometrical features with a known paleowater depth. This approach is similar to the "pinning points" of Franseen and Goldstein (1993) and Goldstein and Franseen (1995) who used a late Miocene outcrop example to reconstruct the relative sea-level
history from lithofacies relationships and sedimentary structures. In this study, I attempt to define anchor point directly from diagnostic stratal geometries (Fig. II.26).

One of the most reliable anchor points is the shelf-slope break in a forced regressive sequence (Handford and Loucks, 1993; Schlager, 1993). The minimum magnitude of falling sea level can be calculated on a 2-D cross section by measuring the vertical difference between the slope breaks in the regressive sequences. Erosional coastal cliffs also serve as good indicators of a sea level fall. The vertical component of the cliff’s face provides a direct measurement of the sea level fall. A potential pitfall lies in confusing the cliffs formed by subaerial erosion with submarine slump and slide scars.

Lowstand prograding wedges or falling stage systems tract (FSST) prograding wedges are good indicators of sea level lowering (Schlager, 1993). Potential errors include misinterpretation of mass wasting deposits such as slumps and slides deposited on the platform slope as FSST prograding wedges. Lowstand prograding wedges are not direct indicators of the sea level fall magnitude. If it can be demonstrated that the deposition of the most shoreward extend of the FSST prograding wedge occurred under coastal conditions, then the past sea level position may be inferred from the first coastal onlap. Two main pitfalls exist in this particular approach. First, if the vertical distance between the previous sea level position and the first coastal onlap is measured in 2-D space, such as on a single seismic line, the magnitude may appear exaggerated if the section cuts the wedge at an oblique angle to the depositional dip. The second pitfall is specific to measurement from seismic sections, where the thickness of the most
Figure II.26. Interpretation of relative sea level position based upon the diagnostic stratal geometries (anchor points).
shoreward extent of FSST wedge may fall below the seismic resolution. This would lead to an overestimation of the sea level fall magnitude.

The upper surface of the flat-top carbonate banks and reef marginal buildups in many cases may serve as a reliable anchor point. While it has been shown that some modern flat-top banks may produce large quantities of sediments even at significant depths (e.g. 25-35 m, northern Nicaragua Rise, Glaser and Droessler, 1991), the majority of flat top banks are shallow. The active (non-drowned) flat-top bank typically maintains its top at or near sea level in 5-20 m of water.

Reef margins are characteristic for rimmed carbonate platforms. In modern settings certain reef-forming biota, such as scleractinian coral Acropora palmata, serve as a reliable indicator of water depth because its habitat is confined to the water depth of 5 m (Fairbanks, 1989). The paleowater depth of the reef margins in the geological record, however, varies significantly. The examples include the Permian of West Texas where the shelf margin reef is interpreted to have existed in 40 to 60 meter of water (Tinker, 1998). Other examples include the late Miocene Majorca prograding complex where red-algae buildups formed the break in slope at the depth of 30 to 70 meters (Pomar, 1999). Still, the measurement of the vertical component of backstepping or aggrading carbonate margin buildups and flat-top banks produced by the same biota would yield a reliable amplitude of the relative sea level rise.

The use of coastal onlap to estimate the magnitude of sea level rise in carbonate environments is complicated by the fact that during a significant sea level rise the entire platform is flooded. In this case, no coastal onlap exists because there is no coastline.
The minimum magnitude of sea level rise can be measured by taking the amount of vertical aggradation of the carbonate bank.

**Application of the anchor point system to the middle Miocene of the Maldives**

Relative sea level position was estimated from three seismic profiles imaging prograding Complex I. The results of the reconstruction, as well as anchor point criteria, are summarized in Table II.2.

The flat top of the early Miocene bank was chosen as the starting anchor point. It was assumed that the top of the bank was at sea level prior to the initiation of bank margin progradation. SARPs consistently demonstrate the basinward shift in onlap and were interpreted as shallow-water carbonates deposited during the falling stage systems tract (FSST prograding wedges). The shoreward extension of these packages is assumed to indicate the position of the relative sea level. It is considered unlikely that these sediments represent gravity-flow or reworked deposits because: 1) there is no evidence of detachment surfaces or chaotic reflectors indicating slumping; 2) SAPRs commonly contain force-regressive geometries; 3) SARPs are clinoform-shaped packages with their thickest parts commonly positioned on the steepest part of the slope. If these were gravity-flow deposits they would be likely thickest at the toe of the slope where the gravity gradient is the smallest; and 4) SARPs are composed of continuous oblique to sigmoidal reflectors indicating gradual, not episodic deposition.
The Sequence 1 SARP forms a wedge attached to the slope of the flat-top bank (Fig. II.9B). The determination of relative sea level position in Sequence 1 is obscured by the local gravitational slide feature described previously. Seismic interpretation indicates however that the sea level dropped below the level of the bank top. The deposition of Sequence 1 WARP occurred during the subsequent relative sea level rise and sea level highstand. The vertical component of Sequence 2 SARP downward shift accounts for -87.5 m on line E130 (Table II.2). The vertical component of the Sequence 2 SARP downward shift measured on line E120 is smaller (-77.5 m) which indicates an oblique cutting of the SARP 2 by line E130. SARP 2 is not present on line E140 (Fig. II.19). The strike cross section of line N060 (Fig. II.22) shows that SARP 2 pinches out to the south. The subsequent flooding represented by LARP 2 onlap gives a vertical aggradation value of 70 m.

This approach is followed to measure the amount of downstepping and aggradation for all of the sequences. In Sequence 4, a more detailed relative sea level history can be reconstructed. In this case of a three-step forced regression (Line E130, Fig. II.17), the shelf-slope breaks serve as anchor points (Fig. II.26). The vertical distance between the Sequence 4 relict shelf breaks is 72.5, 25, and 25 m respectively (Table II.2). The three-step geometry is not present on line E120 but an erosional cliff is clearly displayed (Fig. II.18).

Sequence 5 is characterized by massive gravity-flow deposits accumulated at the toe of the slope. Debris-flow deposits, and the onlapping apron of sediments covering them are not diagnostic of the paleowater depth estimate. On Line E140 (Fig. II.19),
SARP 5 is shifted in the basinward direction relatively to the offlap break of sequence boundary E, implying a large magnitude sea level fall. The onlapping reflector of WARP 5 indicates significant flooding.

Late Miocene sequences 6 and 7 are characterized by progressive onlap indicating filling of the basin. The nature of this onlap is likely to be marine and is not diagnostic for a sea-level position. No reliable anchor points are identified within these sequences because their reflectors extend over the bank top, and the new bank margin is not imaged to the Shell seismic lines. Only the amount of vertical aggradation may be measured, which gives an estimate of the minimum relative sea level rise.

Because of the oblique cutting through the sedimentary package by a 2-D cross section, the magnitudes of sea level falls and rises may be under- or overestimated. The values measured from three seismic profiles should be stacked into a composite record (Table II.2; Fig. II.27). Measurements based on the direct indicators of sea level (anchor points) are given priority over the measurements based upon the ambiguous interpretations (e.g. inferred coastal onlap).

The middle Miocene prograding sequences in the Maldives have been characterized by significant progradation. In prograding Complex I, the bank margin prograded for 11.5 km in approximately 5 Ma. The amplitudes of the middle Miocene flooding events, however, were not large enough to reach the mark of the early Miocene highstand.

The magnitudes of the relative sea level events have not been corrected for compaction or isostatic loading. I am, at this point, reluctant to manipulate the data
Figure II.27. Reconstructed middle Miocene relative sea-level record from the prograding Complex I in the Maldives. The values are not corrected for compaction. Sequence boundaries correspond to the inflection points. The graph shows five high-amplitude sea-level cycles, however, middle Miocene relative sea level remained lower than sea level position at the end of the early Miocene. Because the exact timing of the sequence formation is not known, it was assumed that all sequences took the same amount of time to deposit. The shape of the relative sea level curve will change once the exact timing of sequences is known but the magnitudes of the events should remain the same.
because the physical properties of the sediments are not known. I would like, however, to acknowledge two processes that may have had an important effect on the reconstruction of relative sea level from stratal geometries. First, postdepositional tectonic tilting of the sequences may result from the rotation of a tectonic block after the deposition of sediments. In the case of tilting towards the direction of progradation, the values of the measured sea level magnitudes would be exaggerated. Shell seismic profiles do not provide direct evidence of postdepositional rotation of the strata in the area of the prograding Complex I. A single normal fault displacing the prograding sequences is observed on lines E130 and E140 (Figs. II.9B and II.19), however, the magnitude of the fault throw is not significant. Absence of large faults and the nearly flat topography of the underlying late Oligocene carbonate platform speak for the likelihood of insignificant postdepositional tilting of the strata.

The second important process that affects stratal geometries is compaction-driven subsidence. Hunt et al. (1996) and Hunt and Fitchen (1999) emphasize the role of differential compaction in controlling sequence development. Subsidence caused by differential compaction played an important role in the deposition of sequences in the Maldives. The structure of the pre-middle Miocene sediments consisted of backstepping and aggrading Eocene-early Miocene shallow-water carbonate platforms separated by deeper seaways (Fig. II.24). Although the seaways were entirely filled with sediment by the end of the early Miocene, the soft periplatform carbonate muds deposited in the seaways compacted better than neritic platform sediments. The large part of the prograding Complex I is located above the Oligocene platform and experienced uniform
compaction-driven subsidence. The toes of the clinoforms, however, are positioned above the filled seaway with better-compacted mud. Greater compaction at the clinoform toesets resulted in the development of the sagging stratal geometry (Fig. II.24). The stratal geometries of prograding complex III were subject to even greater compaction-related tilting because of its position above the platform-basin transition (Fig. II.24).
A CASE FOR EUSTASY?

Scale of prograding sequences in the Maldives: Regional or local?

The coverage of the Maldives basin by a dense seismic grid permits a reliable regional correlation of sequences. Prograding complexes I, II and III are located as much as 150 km apart from each other and a difference in the age, number, and expression of sequences would be expected if their formation was indeed governed by a local-scale mechanism such as local subsidence or episodic changes in sediment supply. Our interpretation of the Shell seismic data demonstrated that the five sequences defined in the different prograding complexes are, in fact, synchronous. Figure II.23 shows a cross section through the prograding Complex II. The number of sequences, the timing of their formation, and their geometries are same as in the prograding Complex I (Fig. II.9B). The similarities are extremely important as they indicate that the middle Miocene progradation in the Maldives was not a local phenomenon but rather a basin-wide process. The scale and synchronism of the progradation indicates that the Maldives carbonate system responded to a regional-scale process, which likely included both regional subsidence and eustatic sea-level changes.
Global stratigraphic signature

Thick prograding sedimentary sequences of the middle Miocene age have been described in many basins around the world, including offshore New Jersey (Greenlee and Moore, 1988; Miller et al., 1996; Fulthorpe and Austin, 1988), the Great Bahama Bank (Eberli and Ginsburg, 1989), offshore Alabama (Greenlee and Moore, 1988; Bartek et al., 1991), Bali-Flores sea (Tyrell and Davis, 1989; Bartek et al., 1991), and the Mut basin of Turkey (Broucke et al., 1999, Bassant, 1999). Bartek et al. (1991) argued for the global stratigraphic signature for the Neogene, where early Miocene flooding and aggradation were followed by successive episodes of major progradation in the middle Miocene and late Miocene aggradation. Bartek et al. (1991) suggested that the mechanism responsible for this global signature was the waxing and waning of the Antarctica ice sheets.

In the Maldives, the early-middle Miocene transition is characterized by a turn around from platform margin backstepping in late Oligocene-early Miocene and flat-top bank aggradation in the latest early Miocene to the basin-wide bank margin progradation in the middle Miocene.
Sea level proxies

Benthic foraminifera oxygen isotope data are routinely used as a proxy for the amount of continental ice in the Neogene. The $\delta^{18}$O values reflect continental ice volume because ice preferentially sequesters light oxygen isotopes. The continental ice volume, in its turn, is a proxy of sea level, with periods of rising sea level corresponding to the melting of ice sheets, and periods of falling sea level corresponding to ice sheet growth. Other factors influencing $\delta^{18}$O values are sea water temperature and salinity. Bottom-dwelling (benthic) calcite-secreting organisms are less affected by variations in temperature and salinity occurring in the water column, compared to planktonic organisms, and are thus considered more useful for sea-level studies.

Oxygen isotope data from ODP sites 563 (western North Atlantic), 608 (eastern North Atlantic), and 747 (Indian sector, Southern Ocean) compiled by Wright and Miller (1994) showed a step-wise increase in $\delta^{18}$O values in the second half of the middle Miocene interval. Recent compilations of the benthic foraminifera isotope data show light ($\sim 1\%$) $\delta^{18}$O values at the early-middle Miocene transition (Zachos, 1998, Fig. II.28). The middle Miocene is characterized by a step-wise increase of $\delta^{18}$O values. At the end of the middle Miocene, the $\delta^{18}$O values become as heavier indicating 1.3-1.4% increase. This significant increase implies the growth of continental ice and/or cooling of ocean waters. Using the Mg/Ca paleothermometry method, Lear et al. (2000) suggested that most
Figure II.28. Comparison between the middle Miocene relative sea level record from the Maldives, benthic foraminifera oxygen isotope data (Zachos, 1998), and global sea level curve of Haq et al. (1987). All records are converted to the absolute age scale of Berggren et al. (1995).
(~85%) of the benthic δ¹⁸O increase in the late Middle Miocene (~14 Ma) can be attributed to an increase in the continental ice volume, and the rest of the increase is caused by cooling of the bottom waters. The middle Miocene oxygen isotope record therefore indicates an overall lower sea level in the middle Miocene due to the removal and storage of part of the oceanic water masses in the form of continental ice.

The global eustatic curve of Haq et al. (1987) was constructed from the compilation of coastal onlap charts from sedimentary basins around the world. On the chart, the early part of the middle Miocene (Langhian) is characterized by two sharp, short-term (<1 Ma) sea level falls with amplitudes of 90 m (Fig. II.28, Bur 5/Lan1 and Lan 2/Ser 1 events of Hardenbol et al., 1998). The sea level then rebounded to its previous position. The second part of the Middle Miocene (Serravallian) shows a large three-step sea level fall (Fig. II.28), culminating at the middle-late Miocene boundary. The two falls in Serravallian (Ser-2 and Ser-3 of Hardenbol et al., 1998), each followed by a low-magnitude flooding event, have amplitudes of 50 and 60 m, respectively. The last segment of the fall (Ser 4/Tor 1 of Hardenbol et al., 1998) is the most dramatic and has a magnitude of 140 m. In the late Miocene, the curve shows flooding which culminates in the latest Miocene.

Although plausible mathematical correlation of the timing and, in particular, magnitudes of eustatic events inferred from the benthic oxygen isotope data and the global sea level chart of Haq et al. (1987) is doubtful, both records indicate a step-wise sea level fall in the middle Miocene. This global record agrees with the relative sea-level
record reconstructed from the prograding sequences in the Maldives in this study (Fig. II.28). The results of seismic interpretation based on anchor points indicate five sea-level cycles of significant amplitude (50-100m). Flooding in each cycle is also significant, but relative sea level appears to stay below the late early Miocene position.

Middle Miocene Climatic and Oceanographic Record

The middle Miocene time was characterized by significant changes in the Earth's climatic and oceanographic conditions. An increase in heavy oxygen isotope values in the middle Miocene was interpreted by some scientists (Shackleton and Kennett, 1975; Savin et al., 1975; Flower and Kennett, 1994) to be related to the initiation of the Antarctica ice sheet. Flower and Kennett (1994) suggested that the middle Miocene interval marked a major change in the Cenozoic climatic evolution. They claimed that the early stage of this interval (~16 to 14.8 Ma) was characterized by major short-lived changes in the ice sheet volume, sea level, and global climate. The later stage (~14.8-12.9 Ma) included major East Antarctic Ice Sheet (EAIS) growth and associated Antarctic cooling, large sea level fluctuations followed by a large magnitude sea level fall, and important changes in deep water circulation and production.

Abreu and Anderson (1998) argued that the East Antarctica Ice Sheet had existed since the middle Eocene. They also provided evidence that the West Antarctica Ice Sheet (WAIS) was formed in the early-middle Miocene, and its early evolution included repeated episodes of advance and retreat across the continental shelves. Miller et al.
(1996) compared the timing of the middle Miocene sequence boundaries identified on the New Jersey coastal plain (ODP Leg 150X) and continental slope (ODP Leg 150) with the ages of the $\delta^{18}$O increases. They concluded that the ages of the coastal plain unconformities and slope seismic reflectors matched the ages of the $\delta^{18}$O increases, which were interpreted as sea level falls.

The Glacio-Eustatic Scenario

The interpretation of seismic and well data in the Maldives for this study showed that carbonate bank margin progradation started in the early middle Miocene. Five middle Miocene prograding sequences represent complete sea level cycles. The reconstructed sea-level record, not corrected for tectonic subsidence or sediment and water loading, is in agreement with the global benthic foraminifera oxygen isotope data which implies lower global sea level in the middle Miocene. I suggest that the episodic growth and melting of the Antarctic ice sheet in the middle Miocene time was the primary cause of the basin-wide bank margin progradation in the Maldives. Each prograding sequence represents a complete sea level cycle resulting from the ice sheet growth and a subsequent melting pulse.

The large-magnitude sea level fall in the latest middle Miocene in the Maldives may represent the same eustatic sea level fall that was recorded in the carbonate platform on the Marion Plateau in northeast Australia. Based on seismic interpretation, Pigram et
al. (1992) concluded that a 180-m sea level fall caused the exposure of the shallow water carbonate platform and reestablishment of carbonate production at a greater depth on Marion Plateau in the late Miocene.
CONCLUSIONS

Interpretation of seismic and well data in the Maldives shows that basin-wide carbonate bank margin progradation occurred in the middle Miocene. Five prograding sequences recorded carbonate bank response to relative sea level fluctuations. Their prograding geometries are interpreted as representing five complete sea-level cycles. The cycles are likely to be 0.5-1 million years in duration (third order cycles). The last two sequences recorded sea level falls of significant magnitude and are associated with the accumulation of massive debris-flow deposits.

An attempt to reconstruction relative sea level history and quantify the magnitudes of the middle Miocene relative sea level changes in the Maldives was based on the system of “anchor points”, or paleo-water depth diagnostic geometries. The reconstructed relative sea-level record shows a good similarity with the global sea level record predicted from the benthic foraminifera oxygen isotopes. The initiation and evolution of progradation in the middle Miocene in the Maldives is proposed to be related to the fall of global sea level caused by the systematic growth of the Antarctica ice sheet. The accretion of ice was not continuous and was interrupted by periods of melting resulting in periodic flooding events.

No attempts were made to correct relative sea level magnitudes for subsidence, compaction and isostatic loading because of the absence of information on the physical properties of the sediments forming individual sequences and limited knowledge of
paleowater depths. A multi-well transect with continuous core recovery similar to the New Jersey Margin (ODP Legs 150, 150X, 174) and the Bahamas Transect (ODP Leg 166) is required to address these issues. It is clear, however, that the middle Miocene sequences of the Maldives contain one of the best sea level records available.
GENERAL CONCLUSIONS

The 50-Ma-long evolution of the Maldivian carbonate system may be divided into two main stages. During the first stage (Eocene to early Oligocene), the basement structure played the dominant role on the geographic distribution and development of the shallow-water carbonate production. In contrast, relative sea-level fluctuations controlled the evolution of the carbonate platform during the second stage (late Oligocene to present).

Shallow marine carbonate production was established in the Maldives in the early Eocene on top of faulted volcanic basement. Two NNE-SSW grabens formed troughs that served as a series of intra-platform seaways. In the Eocene and early Oligocene, the carbonate banks rooted on the topographic highs aggraded and backstepped in response to an overall sea-level transgression driven mainly by tectonic subsidence (Fig. III.1). The movement of the graben faults ceased by the end of the early Oligocene, at which time the grabens were substantially filled with the fault scarp material and pelagic sediments.

At the beginning of the late Oligocene (~28.5 Ma), a significant (>50 m?) sea-level fall exposed the bank tops and explained the observed in ARI-1 well decrease of the water depth in the seaways (Fig. III.2A). This sea-level fall corresponds to a significant increase of deep-water benthic foraminifera oxygen isotope values and is interpreted to
Figure III.1. Schematic west-east cross section through the Maldives carbonate platform. Strongly modified from van Gils and Rubbens (1992).
ON THROUGH THE MALDIVE PLATFORM
Figure III.2. Schematic evolution of stratal geometries of the Maldives carbonate platform in the early/late Oligocene - middle Miocene.
be a eustatic in origin. Shallow carbonate production resumed in the late Oligocene. Carbonate banks backstepped in response to a relative sea-level rise (Figs. III.1 and III.2B). The late Oligocene banks developed a series of elevated rimmed margins that separated the deep internal basins from the bank interior.

The late Oligocene-early Miocene transition was characterized by a continued transgression that was expressed by the backstepping of the bank margins and the partial drowning of the bank tops. In the large central Gaha-Male bank, the late Oligocene platform interior, characterized by numerous patch reefs, drowned while the platform rim continued to aggrade, creating a characteristic “empty bucket” geometry (Fig. III.2C).

The backstepping of carbonate platforms in the early Miocene was accompanied by the infilling of the existing seaways with periplatform and pelagic sediments (Fig. III.1). By this time, the influence of the basement topography on shallow-water carbonate production was negligible. The overall backstepping resulted in the establishment of flat-top elongated and narrow carbonate banks on the periphery of the basin (Fig. III.2D). The late early Miocene banks were characterized by significant vertical aggradation (Fig. III.1.).

At the beginning of the middle Miocene, sea level fell and probably exposed the flat-top banks on the periphery the carbonate system. In the middle Miocene, the bank margins prograded for significant distances (Figs. III.1 and III.2E). Five prograding sequences were defined in the individual prograding complexes. Each prograding sequence formed in response to a complete sea-level cycle. The position of relative sea
level was lower than at the end of the early Miocene and the tops of the early Miocene bank were likely exposed.

In the late Miocene and early Pliocene, the relative sea-level rise caused the flooding of the exposed early Miocene bank tops, which aggraded vertically and prograded locally (Fig. III.1). The early-late Pliocene boundary is recorded as a systematic downward shift of base level. During the late Pliocene, the carbonate production was limited to a narrow zone in front of the partially exposed early Pliocene bank top. In the Quaternary, the periodic flooding and exposure of the bank tops caused by the waxing and waning of continental ice caps led to the transformation from a flat-top bank to the present-day atoll morphology.

The reconstructed late Oligocene – middle Miocene relative sea level history for the Maldives shows good similarity with the eustatic record predicted from the deep-water benthic foraminifera oxygen isotope compilation (Zachos, 1998). Despite the fact that the exact amplitudes of the sea-level events are difficult to establish and correlate at this time, the general sea-level trends based upon the oxygen-isotope proxy are observed in the depositional signature of the Maldives. This implies that sea-level events captured in the Maldivian record are global, or have a dominant eustatic component.

A difference between the two records exists at the Oligocene-Miocene transition. The interpretation of the Maldivian data suggests a continuous transgression in the late Oligocene-early Miocene while the isotope data implies a lower sea level in the early part of the early Miocene. This discrepancy may be explained by the effect of deep water cooling on the increase of the oxygen isotope values, not associated with the growth of

The future research in the Maldives should be the ground-truthing of the seismic interpretations developed in my dissertation. Continuous sediment recovery by drilling will lead to a more precise estimate of the timing and magnitudes of sea-level events recorded in the Maldives based on refined stratigraphy, paleobathymetry and lithology of the individual sequences. This could be achieved by conducting a drilling transect in the Maldives similar to those drilled on the New Jersey Margin (ODP Leg 150, Mountain et al., 1996) and in the Bahamas (ODP Leg 166, Eberli et al., 1997). A proposal to drill such transect entitled “Timing and Amplitudes of Oligocene/Miocene sea level fluctuations in the Inner Sea of the Maldives Archipelago: an intra-oceanic carbonate system (equatorial Indian Ocean)” was submitted to ODP and is presently under review (Droxler and Belopolsky, ODP proposal #514-Full4). The drilling of this transect would be a logical continuation of the research in the Maldives carbonate platform – an extensive archive of Earth’s history for the last 50 million years.
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