ORIGIN OF DOLOMITE IN THE ARAB-D RESERVOIR FROM THE GHAWAR FIELD, SAUDI ARABIA: EVIDENCE FROM PETROGRAPHIC AND GEOCHEMICAL CONSTRAINTS

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ABSTRACT: A significant proportion of oil production from the Kimmeridgian Arab-D strata in the Ghawar field, Saudi Arabia originates from dolomitized rocks. Stratigraphic, petrographic, and geochemical data suggest that at least four episodes of dolomitization affected these sediments. The lower portion of the Arab-D, Zone 3, is only partially dolomitized, with the dolomite frequently being associated with firm-grounds. We propose that these dolomites were formed on an outer-ramp setting with a maximum water depth of 50 m, during a period of nondeposition, with the dolomitization process being promoted by the oxidation of organic material and the diffusion of Mg^{2+} from the overlying seawater. The dolomites in Zone 2B are geochemically distinct compared to those in Zone 3 in that they have relatively positive oxygen isotope compositions (-1 to -2% compared to -6.5%). The relatively positive oxygen isotope composition and the geochemical similarity of Zone 2B to the dolomites in Zone 1, which are intimately associated with the overlying evaporites, has led us to conclude that the Zone 2 dolomites probably formed by the reflux of hypersaline fluids through the sediments. These hypersaline fluids bypassed Zone 2A by moving through the grain-dominated sediments. Early cementation and dolomite formation made these units more susceptible to later fracturing that affected the entire Arab-D formation. This fracturing allowed higher-temperature fluids to leach the dolostones, thereby removing any remaining calcite and partially resetting the oxygen isotope composition of some of the dolomites. As a result of this later dolomitization event, rocks that were only partially dolomitized were leached, creating units with extremely high permeability and porosity (super-k intervals). Dolomites in the lower Zone 3 were recrystallized during burial by the normal geothermal gradient, leading to the present negative oxygen isotope values. Zone 1 dolomites are petrographically distinct from Zone 2 dolomite in that they are mimetic and fabric preserving, although they are geochemically similar. This mimetic style of dolomitization occurs immediately adjacent to the overlying anhydrite and is interpreted to have formed very shortly after deposition from hypersaline brines.

INTRODUCTION

The Arab-D reservoir in the Ghawar Field of Saudi Arabia is the world's largest oil field producing approximately 5 million barrels a day (Durham 2005) (Fig. 1). The greater part of the production is derived from grain-dominated limestones and although the dolostones in the Arab-D in general have rather poor reservoir characteristics, the presence of dolomite has a pronounced impact on reservoir quality in a number of ways. In some places, dolostones are responsible for producing permeability barriers, whereas in other places they are associated with zones of very high production (Meyer et al. 2000). These zones of very high flow have been termed "super-k" zones and are defined as intervals that give a flow of oil or water of greater than 500 barrels per day per foot of vertical interval (Meyer et al. 2000). Frequently super-k zones are associated with poor core

recovery, and their presence is inferred from a combination of production data and the diameter of the bore hole as measured using the caliper log. As a result of the importance of dolomite in the formation of the super-k intervals within the Arab-D, an understanding of the paragenesis of these zones is important not only in predicting super-k formation within the Arab-D but also as an analogue for understanding other dolomitized reservoirs with high permeability and porosity.

Previous characterization work of dolostones from the Arab-D in the Ghawar field (Cantrell et al. 2004) has identified three types of dolomite with distinctive petrographic, stratigraphic, and geochemical characteristics: fabric-preserving, non-fabric-preserving, and baroque dolomite. Fabric-preserving dolomite is very finely crystalline in which petrographic details of the original limestone fabric are usually well preserved. Fabric-preserving dolomite contains low concentrations of Fe (average 247 ppm Fe) and relatively positive oxygen isotope values (average $\delta^{18}O = -2.08\%$). Beds of fabric preserving dolomite typically occur as thin, sheet-like or stratigraphic layers in the uppermost interval of the Arab-D that are always intimately associated with the transition from carbonates below to anhydrite above. Non-fabric preserving dolomites are medium crystalline, non-baroque dolomite in which all traces of the original sedimentary fabric have been obliterated. The non-fabric-preserving dolomite has a relatively low concentration of Fe (152 ppm) and moderately negative oxygen isotope values (average $\delta^{18}O = -2.66\%$). This dolomite also typically occurs as stratigraphic beds, although it is present throughout the Arab-D and is not restricted to the carbonate-evaporite transition in the uppermost portion of the Arab-D. The third type, baroque, is a coarsely crystalline dolomite with "saddle-shaped" crystals displaying undulose extinction in thin section. Geochemically, this dolomite is distinct from the fabric-preserving and non-fabric-preserving dolomite types, in that it typically contains high concentrations of Fe (31,500 ppm) and fairly negative oxygen isotope values (average $\delta^{18}O = -7.37\%$). Baroque dolomite is rare in the reservoir and appears to be limited to areas that contain abnormally thick dolomites; in extreme cases, it is vertically pervasive. In contrast to the approach utilized by Cantrell et al. (2004), which examined only the selected dolomitized intervals from a large number of wells scattered throughout the Ghawar field, we have examined the petrography and geochemistry of three wells from the Haradh region in detail (Fig. 1).

Background

During Middle Jurassic time the Arabian platform was situated at about 10° S and faced the Tethys Ocean which lay to the present-day east and northeast. By that time a large intrashelf depression, called the Lurestan Basin, had developed in the Arabian Platform (Fig. 2A). From the middle Jurassic to the Turonian (Cretaceous) the Lurestan Basin became differentiated into the Gotnia Basin (Basrah Basin) and the Rub'al Khali Basin (Murriss 1980). A structural arch (The Rimthan Arch), which separates the two basins, stood high relative to the basins on either side, as suggested by the thick grainstones deposited there during Kimmeridgian time (Wilson 1975). This grainstone belt and the northern side of the Rimthan Arch

FIG. 1.—Map showing location of the Ghawar field and the location of wells A, B, and C used in this study. The wells are located in the Haradh region of the Ghawar field. Also shown are the names of other subdivisions in the Ghawar field.

probably formed a fairly well defined shelf margin that faced the Gotnia Basin to the present-day north. The shelf-margin passed toward the interior of the Arabian platform (present-day south) and into the Rub'al Khali, where Ghawar field is located. The southward transition from the Rimthan Arch and shelf-margin into the Rub'al Khali Basin was gradual. This is indicated by (1) a southward reduction in the volume of grainstones in the Arab-D carbonates and (2) a southward thinning of the Arab-D carbonates and thickening of the evaporite section (Mitchell et al. 1988). Mitchell et al. (1988) explained these relationships as a record of a lateral facies change and shallowing from shoal-water carbonates in the north to coeval sabkha evaporites in the south. More recent work has suggested that the evaporites are not coeval with the carbonates and that the evaporites record shallow subaqueous, or salina and desiccated salina deposition rather than supratidal sabkha deposits (Handford et al. 2002). This is a fundamental paradigm shift from previous hypotheses because it proposes that evaporite deposition occurred later than the carbonates and that the evaporites filled the remaining accommodation space in the Rub'al Khali Basin during an evaporative drawdown as part of a lowstand systems tract.

Arab-D lithofacies classification schemes are based largely on the work of Mitchell et al. (1988) (Fig. 2B). Internally within the reservoir, deposition of cyclic shoaling-upward packages are identified using a succession

of porosity breaks in porosity well-log patterns. Operationally, these porosity breaks are used to differentiate the six major reservoir zones depicted in Figures 2B and 3; these zones, from the top of the reservoir down, are zones 1, 2A, 2B, 3A, 3B, and 4. With the exception of the Zone 1 and 2A boundary, these zone boundaries are tentatively identified as timelines (Mitchell et al. 1988; Meyer et al. 2000). An idealized section through the Arab-D is shown in Figure 3. The top of the Arab-D (Zone 1) is characterized by thin subtidal to intertidal carbonates with evaporites. Carbonates in this unit are frequently completely dolomitized. Underlying the evaporites is a series of grainstones and packstones (zones 2A and 2B). The grainstones in Zone 2A are frequently undolomitized while the grainstones and packstones in Zone 2B are locally pervasively dolomitized and compose a large proportion of the reservoir in the Haradh region of the Ghawar Field. Both skeletal and non skeletal grains are common, with the skeletal material being composed of benthic foraminifera, dasycladacean green algae, corals, red algae, bivalves, and stromatoporoids. The stromatoporoid *Cladocoropsis* is particularly important in zones 2A and 2B, where it is frequently associated with zones of extremely high porosity and permeability (super-k zones; Meyer et al. 2000). Zone 3 is composed mainly of micritic limestones interbedded with grain-supported carbonates (Fig. 4).

FIG. 2.—**A**) Paleogeography of the Ghawar field during the Late Jurassic; the Luestan Basin is composed of the Gotnia and Rub'al Khali Basins and is not shown for clarity. The carbonates have been interpreted as being deposited on the transition from a shallow carbonate shelf to an intrashelf basin (Handford et al. 2002). **B**) Schematic of the generalized terminology used to describe the Arab-D (Cantrell et al. 2001).

FIG. 3.—General schematic of the stratigraphy and sedimentology used to describe the Arab-D, based on Mitchell et al. (1988). The core shown is an idealized composite, and the depths are shown for demonstrative purposes only. Zone 1 can have interpolated layers of calcite, dolomite, and evaporite.

The rocks are partially dolomitized, although the amount can vary substantially over short distances. Increased concentrations of dolomite occur in association with marine nondepositional surfaces (NDS) and in associated burrow fills.

METHODS

This study concentrated on three wells from the Haradh region of the Ghawar field (Fig. 1) and analyzed bulk rocks and chemically isolated

FIG. 4.—Detailed sedimentological description and mineralogy as estimated from hand-sample description in Zone 3 from Well C. Note that the increases in the amount of dolomite occur within this zone immediately below nondepositional surfaces (NDS). Such increases occur below NDS at 7307', 7315', 7320', 7331', 7336', 7341', and 7388' (2227, 2229.6, 2231.1, 2234.5, 2236, 2237.5, and 2251.9 m). Surfaces at 7328', 7344', 7347', and 7349' (2233.5, 2238.5, 2239.4, and 2240 m) show no increases in concentration of dolomite immediately below the NDS. An increase in the concentration of dolomite at 7380' (2249.4 m) does not appear to be associated with any recognized NDS.

dolomites from samples at intervals of between 0.3 to 1 m from zones 3, 2, and 1 of the Arab-D (Fig. 3). Approximately 850 samples were analyzed. Standard petrographic analyses of thin sections were performed to characterize the lithofacies and paragenetic sequence. Petrographic analyses include the description of non dolomitized constituents, amount of dolomite, dolomite type, porosity type, and diagenetic features.

In preparation for wet chemical analyses and X-ray diffraction (XRD), all materials were ground to finer than 63 μm . For XRD analyses, powdered samples were smear-mounted on glass slides. Peak areas for aragonite, calcite, and dolomite were determined using a Scintag XDS-2000 diffractometer. Mineral concentrations were calculated from peak area ratios assuming that each sample was composed only of calcite, aragonite, and dolomite. Peak area ratios were calibrated and the concentrations calculated using calibration curves prepared from results using a series of pure mineral standards (verified by XRD analysis). Duplicate analyses indicate a reproducibility of $\pm 3\%$. In addition to X-ray diffraction, estimates of dolomite abundance were made by examining pore plugs taken at intervals of 0.25 m. In order to isolate the dolomite, selected sieved samples with greater than 5% dolomite were treated with buffered acetic acid for a period of 2 hours (Swart and Melim 2000). This procedure selectively leaches the less stable minerals, leaving the dolomite behind. After the leaching procedure the samples were again subjected to X-ray analysis in order to check that the samples consisted entirely of dolomite. If the samples still contained some nondolomite carbonate minerals, the leaching procedure was repeated. This method not only removes nondolomite carbonate minerals but also has the effect of removing adhering contamination from the samples. Treatment of a standard dolomite in this manner has been shown not to influence significantly the geochemistry of the sample, although we cannot rule out the possibility that the treatment preferentially removed some of the dolomite itself.

For $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ analyses, all samples were reacted for 10 minutes using the common acid bath method at 90°C and the CO_2 produced analyzed using a Finnigan-MAT 251. Reaction was carried out for a period of 10 m at 90°C. Standard isobaric corrections were applied, but no correction was applied for the differences in the fractionation of $\delta^{18}\text{O}$ as a result of the dissolution of dolomite and calcite by phosphoric acid (Land 1980; Vahrenkamp and Swart 1994). Data for both C and O isotopes are reported relative to Vienna Pee Dee Belemnite (V-PDB) using the conventional notation.

The strontium concentrations of the pure dolomite separates were determined using atomic absorption (Perkin-Elmer 4500). In this method approximately 100 mg of dolomite separate was dissolved in 10 cm^3 of 10% nitric acid solution, filtered, and the filtrate diluted to 25 cm^3 with deionized water. Corrections were made for the percentage of insoluble residue. Standards were made using specpure CaCO_3 and MgCO_3 (Johnson-Matthey), weighed out in approximately the same concentrations as contained in the samples and dissolved in 4% HNO_3 . These solutions were then spiked with Sr, Mn, and Fe to provide standards with similar intensities and matrix interferences to the analyzed samples. Reproducibility of this method is approximately $\pm 5\%$. The procedure used for concentrating the dolomite samples also removes adhering contamination in the form of iron and manganese oxides. As a result of the procedure of concentrating the dolomites, contributions of Sr, Mn, or Fe leached from insoluble residues were minor.

Porosity and permeability were determined at intervals of 0.5' (15.2 cm) on 1" (2.54 cm) plugs drilled from the core using standard methods (American Petroleum Institute 1998).

RESULTS

Petrography

Zone 3.—Most samples in Zone 3 are partially dolomitized mudstones to wackestones/packstones and contain considerable amounts of preserved calcite (Fig. 5A, B). Although dolomite crystals are euhedral, reaching

about 100 μm in size, visible porosity is low. Burrows and borings are preferentially dolomitized and stylolites surround and cut dolomite rhombs, indicating post-dolomitization chemical compaction. Occasionally molds are observed, especially in the upper part of Well A. In Well A, a 4 m interval of pure sucrosic dolomites is present. In this interval, dolomite crystals reach 200 μm in size, and frequently show corroded zones (Fig. 5C).

Zone 2.—Samples from Zone 2 are coarser-grained than those from Zone 3 with wackestones to grainstones showing varying proportions of euhedral dolomite (Fig. 5D). Crystal sizes are, however, larger than in Zone 3 and increase upwards from about 150 μm , at the base of Zone 2, to 400 μm in the mid-portion before decreasing to less than 100 μm towards the top of Zone 2. This shift in crystal size roughly parallels the proportion of dolomite in the samples. Zone 2B is characterized by partially dolomitized wackestones to packstones, with porosity being dominated by moldic and inter-granular (primary) porosity. Pre-lithification (mechanical) compaction as well as chemical compaction (stylolites) is observed. In Well C, the super-k interval is located at the base of Zone 2B. This interval consists of partially dolomitized grainstones and packstones. Dolomite crystals are generally fabric destructive. In contrast to wells A and B, the super-k in Well C is not associated with a larger than normal bore hole diameter and is situated lower in the stratigraphic succession. In Well A the super-k is inferred from production data and a larger than normal caliper log (Fig. 6). The larger than normal bore hole in this interval in wells A and B suggest extensive leaching. The super-k interval in Well B is composed of sucrosic dolomites (300 μm) with not only good intercrystalline porosity but also locally tight dolomite. Post-dolomitization solution seams are occasionally observed. Above the super-k interval in Well B, tight dolostones with lower porosity are present (Fig. 5E). This interval is characterized by dolomite crystals, euhedral to anhedral in shape. The dolomite crystals vary in size from 120 to 400 μm and show a distinct zonation comprising cloudy centers and clear rims. In some samples, dolomitization seems to be related to fractures (6583.7' and 6587.7') (2006.7 m and 2007.9 m), where subhedral to anhedral dolomite occurs adjacent to fractures in samples otherwise dominated by sucrosic dolomite. Replacive silica is frequently observed in this interval. Porosity is moldic and intergranular and is generally low. Zone 2A is characterized by regions with low concentrations of dolomite (with the exception of a few pure dolomite layers in Well A). This zone consists dominantly of partly dolomitized to nondolomitized wackestones, packstones, and grainstones. Grainstones show low but varying degrees of cementation. Syntaxial calcite overgrowths occur on echinoderm debris and brachiopod shells. Debris of *Cladocoropsis* generally shows cement lining the internal pores. Dolomite rhombs where present are fabric destructive. Porosity is low where primary intergranular pores and molds are absent. Occasionally, evaporite minerals occlude pore space. Micritic rims and pre-lithification (mechanical) compaction, indicated by breakage of ooids, elongated bioclasts and micritic rims, and by pressure indents, represent earliest diagenetic alterations.

Zone 1.—Zone 1 samples consist of low-porosity partially dolomitized mudstone to grainstone as well as pure dolostones. Fabric-preserving (mimetic) dolomite was only observed in this Zone (Fig. 5F). Anhydrite infills a small portion of the porosity and was observed to replace the calcitic matrix. Well A differs from wells B and C by the presence of leaching features such as vugs, molds, and dedolomite.

Difference between Wells.—In addition to the differences in lithofacies between the varying stratigraphic zones, pronounced lateral differences in diagenetic style can be observed between the three wells. The upper part of Well A (zones 1 and 2) is distinguished by pronounced dissolution features such as leaching of the calcitic matrix and components, and corroded dolomite crystals (Fig. 5C). Dissolution of the cores of dolomite rhombs is common throughout zones 1 and 2, creating secondary intracrystalline porosity. In several samples, this process seems related to dissolution seams that might have formed contemporaneously. In these solu-

FIG. 5.—**A)** Partly dolomitized grainstone (7318.7') (2230.7 m) and **B)** mudstone (6571.6') (2003.0 m) from Well C. **C)** Dissolution of dolomite rhombs in a sample that has been completely dolomitized (Well A, 6335.0') (1930.9 m). **D)** Sucrosic dolomite from Well B (6571.6') (2003 m) showing high intercrystalline porosity. **E)** Dense dolomite from Well B (6576.7') (2004.6 m). **F)** Moldic dolograins from Zone 1 (Well C, 7194.8') (2192.9 m). Scale bar = 500 μm .

tion seams, hollow dolomite rhomboids are observed to have collapsed as a response to compaction. Zone 3 of Well A in contrast, is characterized by low visible porosity and the absence of corroded dolomite. Well C differs from the other two wells in the abundance of dolomite, with significant variations in crystal size (50 to 400 μm) and/or shape (euhedral and anhedral). In contrast to Well A, leaching and dolomite dissolution were not observed. High-porosity intervals in Well A are caused by the presence of sucrosic dolomite (super-k interval) and by preserved primary porosity in grainstones (upper part of Zone 2). Well C exhibits the lowest

visible porosity with the exception of the super-k interval in Zone 2 and some partly cemented grainstones just below this interval. Dedolomitization is restricted to the lower part of Zone 3, and is characterized by replacement calcite in the cores of dolomite rhombs.

Mineralogy

The mineralogy as measured using X-ray diffraction is presented in Figures 6, 7, 8, and 9. All geochemical data reside in JSR's digital repository

FIG. 6.—Geochemical parameters from Well A. A) Mineralogy, B) porosity and permeability (permeability is shown by the solid line and porosity by the dashed line), C) oxygen and carbon isotopes of bulk samples, D) well caliper. Zonation of reservoir is shown on the right hand side of the figure and is based on log correlations. Position of the super-k zone is inferred from caliper log and well production data and is shown as horizontal shaded box. In the case of well A the super-k zone was not recovered. No moving correlation was performed (See Figs. 7 and 8) between the percentage dolomite in the bulk sample and the $\delta^{18}\text{O}$ because a large portion of the core was missing as a result of poor recovery.

(see Acknowledgments section). The mineralogical composition varies between 100% dolomite and 100% low-Mg calcite, with Zone 1 being composed of a mixture of anhydrite, calcite, and dolomite. In addition to low-Mg calcite and dolomite, minor amounts of quartz were detected, associated with the super-k zone in Well C. Zone 3 tends to contain lower concentrations of dolomite with only a few intervals that are completely dolomitized. On the basis of estimates from thin sections and bulk samples, higher concentrations of dolomites tend to occur below NDS (nondepositional surfaces) and decrease with increasing depth away from these surfaces. This variation is reflected in the alternations between dolomite and calcite in Zone 3 shown in Figures 6, 7, 8, and 9. Detailed variations in the sedimentology relative to the mineralogy in Zone 3 for Well C is shown in Figure 4. Although increases in dolomite generally occur immediately below NDS, not all NDS show evidence of increased concentrations of dolomite, and there is at least one interval with higher concentrations of dolomite, that is apparently not associated with a NDS. In contrast to Zone 3, Zone 2 is frequently composed of 100% dolomite, although there are still partially dolomitized or nondolomitized intervals, particularly in Zone 2A.

Stable Isotopes

Oxygen.—The stable oxygen isotope composition ($\delta^{18}\text{O}$) of the bulk samples varies from approximately -0.6 to -6.5 ‰ (Figs. 6, 7, 8, 9, and

10). A comparison of the $\delta^{18}\text{O}$ data for zones 1, 2, and 3 are shown in Tables 1 and 2. The range of $\delta^{18}\text{O}$ values presented in this paper are similar to those previously reported for dolomites from the Arab-D of the entire Ghawar field (Cantrell et al. 2004). The most positive $\delta^{18}\text{O}$ values are found within zones 1 and 2, and in wells B and C, where the super-k zones are located. A more detailed example of change in Zone 2B within Well B is shown in Figure 9. There is no statistically significant correlation between the $\delta^{18}\text{O}$ of the bulk sample and the percentage of dolomite (Fig. 10B), although the samples with the highest $\delta^{18}\text{O}$ values always contain nearly 100% dolomite. If the correlation between percentage dolomite and the $\delta^{18}\text{O}$ of the bulk sample is examined relative to the different zones, it is clear that in zones 1 and 2B there is a positive correlation between percentage dolomite and $\delta^{18}\text{O}$ of the bulk samples, while in Zone 2A and 3 this relationship is reversed (Fig. 11). A 10-sample windowed regression coefficient is shown in Figures 7 and 8 for wells B and C; this relationship is not shown for Well A as a result of the large interval of zero recovery in that Well. This type of correlation examines the association between dolomite and $\delta^{18}\text{O}$ for successive intervals containing 10 samples. The implications and significance of these trends are discussed later in the paper.

A large number of dolomite-rich samples from both wells B and C were leached in order to remove all traces of calcite and then reanalyzed for their $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ compositions. The resultant analyses of the dolomite

FIG. 7.—Geochemical parameters from Well B. A) Mineralogy, B) porosity and permeability (permeability is shown by the solid line and porosity by the dashed line), C) oxygen and carbon isotopes of bulk samples, D) correlation between the percent dolomite and the oxygen isotope composition (see text for details). For nine degrees of freedom (the width of the moving window) a value of 0.6 is statistically significant at the 95% confidence limits. This is marked on the figure. Zonation of reservoir is shown on the right hand side of the figure and is based on log correlations. Position of the super-k zone is inferred from caliper log and well production data and is shown as a horizontal shaded box.

Fig. 8.—Geochemical parameters from Well C. A) Mineralogy, B) porosity and permeability (permeability is shown by the solid line and porosity by the dashed line), C) oxygen and carbon isotopes of bulk samples, D) correlation between the percent dolomite and the oxygen isotopic composition (see text for details). For nine degrees of freedom (the width of the moving window) a value of 0.6 is statistically significant at the 95% confidence limits. This is marked on the figure. Zonation of reservoir is shown on the right hand side of the figure and is based on log correlations. Position of the super-k zone is inferred from caliper log and well production data and is shown as a horizontal shaded box.

separates frequently showed little difference from the bulk samples analyzed from the same interval, and a t-test comparing the mean $\delta^{18}\text{O}$ of the samples before and after the isolation procedure showed no statistically significant difference. This suggests that the calcite removed from the samples has a $\delta^{18}\text{O}$ value similar to that of the dolomite in the sample.

Carbon.—The $\delta^{13}\text{C}$ of the bulk samples all showed relatively positive values (Figs. 6, 7, 8, 9, and 10). A comparison of the $\delta^{13}\text{C}$ data for zones 1–3 is shown in Tables 1 and 2. In contrast to $\delta^{18}\text{O}$, there is a positive correlation between the percentage dolomite and the $\delta^{13}\text{C}$ values ($r^2 = 0.51$, $n = 331$, $p > 0.01$, Fig. 10A). Dolomite separates show slightly more positive $\delta^{13}\text{C}$ values than the bulk samples. Overall there is a positive correlation between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ ($r^2 = 0.24$, $n = 331$, $p > 0.01$, Fig. 12).

Trace Elements

Strontium.—The concentration of strontium was measured in the dolomite separates in wells B and C. The mean Sr concentration in Well B was 101 ppm and in Well C 129 ppm. There was no statistically significant difference in the concentration of Sr between the two wells at the 95% confidence limits. The results from these analyses (Figs. 9, 13A, B, D) show that while the concentrations of Sr are generally low, there are occasional values as high as 1000 ppm in some samples. In both wells B and C, samples taken from Zone 1 showed higher values than other zones. Zone 3B also showed slightly higher Sr concentrations compared to zones 2A, 2B, and 3A.

Iron.—The concentration of iron averages approximately 400 ppm in

Fig. 9.—More detailed view of changes in mineralogy, porosity, permeability, $\delta^{13}\text{C}$, $\delta^{18}\text{O}$, and Sr concentration within Zone 2B of Well B (solid shading = dolomite). The super-k zone is indicated by the shaded area. Note that the $\delta^{18}\text{O}$ of the rock, which is composed of almost 100% dolomite decreases in the super-k interval. There are decreases in the $\delta^{18}\text{O}$ in other portions of the core immediately above and below the super-k interval, but these are associated with decreases in the amount of dolomite. The R-value refers to a 10 point moving regression coefficient between % dolomite and $\delta^{18}\text{O}$. The $\Delta^{18}\text{O}$ dolomite-calcite is the difference between the $\delta^{18}\text{O}$ of the dolomite and calcite.

FIG. 10.—**A**) Percentage of dolomite vs. $\delta^{13}\text{C}$, and **B**) percentage of dolomite vs. $\delta^{18}\text{O}$ in bulk samples. The carbon isotope and the concentration of dolomite are positively correlated ($r^2 = 0.51$, $n = 331$, $p > 0.01$). In contrast, the $\delta^{18}\text{O}$ of the bulk samples is poorly correlated with the percentage of dolomite. There appears to be a wide range of oxygen isotope compositions of dolomite, but a rather narrow range of oxygen isotope composition for the LMC component.

both wells with values ranging in Well B from 290 to 1800 ppm and in Well C from 200 to 1400 ppm (Fig. 13B and 13E). No significant trends were observed relative to depth in the well. The concentration of Fe does not show any statistically significant correlation with either the percentage of dolomite, the concentration of Sr, or the oxygen or carbon isotope composition.

Manganese.—The concentration of Mn averages 27 ppm for both wells and shows a statistically positive correlation with Fe ($r = 0.55$, $n = 79$, $p > 0.01$, Fig. 13C, F).

DISCUSSION

Stable Isotopic Data

Oxygen.—The range of $\delta^{18}\text{O}$ values (-1 to -6.5‰) measured in this study is similar to that recently reported from a large number of wells in the Arab-D (Cantrell et al. 2004). However, this study shows $\delta^{18}\text{O}$ differences within individual cores equivalent to the entire range of $\delta^{18}\text{O}$ values previously reported in the Arab-D of the Ghawar field. These differences exist despite apparent similarities in the petrographic types of dolomite. For example, in Well B the interval between 6560 and 6600' (1999.5 and 2011.7 m) possesses non-fabric-preserving sucrosic dolomites, all of which have relatively positive $\delta^{18}\text{O}$ values (-1 to -2‰). In contrast, the dolomites below this interval that are petrographically similar contain much more negative $\delta^{18}\text{O}$ values (-4 to -6.5‰). Hence dolomites that are separated by only 50' (15 m) in depth have a range of $\delta^{18}\text{O}$ values as wide as 6‰. In order to form these two dolomites at the same time, a difference in temperature or fluid composition of about 25°C or 6‰, or some combination would be needed. Using Quaternary dolomite occurrences from the Bahamas as an analogy (Supko 1977; Kaldi and Gidman 1982; Dawans and Swart 1988; Swart and Melim 2000), it would be extremely unusual for dolomites that have a common origin to have such wide range in $\delta^{18}\text{O}$ values (Fig. 12). Therefore, either the $\delta^{18}\text{O}$ values of one or both of the dolomites have been altered by later diagenetic processes, or the dolomites

formed at different times and/or from different fluids. In comparison with the $\delta^{18}\text{O}$ values of modern analogues, even the dolomites in our study with the most positive $\delta^{18}\text{O}$ values are approximately 5‰ more negative than modern Bahamian equivalents. At least part of this difference (1 to 2‰) can be attributed to a more isotopically negative ocean composition during the Jurassic than at the present time (Veizer 1974; Veizer and Hoefs 1976; Lohmann 1987). The remaining 3 to 4‰ represents either an original depositional temperature 12 to 16°C warmer than at present time, or the recrystallization of the dolomite in the presence of isotopically more negative fluids.

In Well B, the super-k zone which accounts for over 70% of the production of the well (Figs. 7, 9), is a 4 ft (1.2 m) interval between 6581 and 6585' (2005.9 and 2007.1 m) located in the middle of Zone 2B, and corresponds with the zone of more positive $\delta^{18}\text{O}$ values. This is similar to Well C, where the super-k zone is slightly thicker (Fig. 8), but is situated at the base of Zone 2B. The super-k is also associated with a bore hole caliper log in wells A and B that indicates a larger than normal bore hole diameter, a sonic and neutron density log that indicates high porosity, and core samples with high porosity and permeability. Petrographic examination of this interval reveals a small amount of quartz and anhydrite, a finding confirmed in Well C by XRD analysis. Within the super-k of wells B and C, the $\delta^{18}\text{O}$ of the bulk rock was positively correlated with the percentage of dolomite (Figs. 7, 8). Such a correlation is normally expected in partially dolomitized rocks as a result of the approximately 3‰ equilibrium enrichment of dolomite relative to LMC (Land 1980). However, if this relationship is examined in the dolomites from the three wells, then it is evident that there is no general correlation between the concentration of dolomite and $\delta^{18}\text{O}$ (Fig. 10). In fact there are some intervals in which there is actually an inverse correlation between these two parameters. Figure 11 illustrates an example of this phenomenon with 15 consecutive samples selected from four portions of Well B representing Zone 1, 2A, 2B, and 3. In Zone 1 there is a positive correlation between $\delta^{18}\text{O}$ and the percentage dolomite. This changes to an inverse correlation in Zone 2A, no correlation

TABLE 1.—Comparison of geochemical parameters from Well B for chemically isolated dolomite (C, O, % dolomite, Sr, Fe, and Mn).

	C ‰	O ‰	% Dolomite	Sr (ppm)	Fe (ppm)	Mn (ppm)
Zone 2A	2.10 (0.51)	-3.19 (0.82)	34 (47)	517 (426)	339	16
Zone 2B	2.45 (0.57)	-2.97 (1.44)	66 (38)	100 (118)	386 (311)	27 (5)
Zone 3A	2.42 (0.35)	-4.49 (0.82)	47 (39)	92 (42)	476 (430)	27 (4)
Zone 3B	2.50 (0.33)	-4.44 (0.78)	45 (37)	98 (49)	451 (155)	29 (4)

Values in brackets represent one standard deviation; nm indicates that no samples were measured.

TABLE 2.—Comparison of geochemical parameters from Well C for chemically isolated dolomite (C, O, % dolomite, Sr, Fe, and Mn).

	C ‰	O ‰	% Dolomite	Sr (ppm)	Fe (ppm)	Mn (ppm)
Zone 2A	2.53 (0.57)	-3.56 (0.46)	41 (41)	427 (273)	717 (354)	25 (3)
Zone 2B	2.40 (0.43)	-4.04 (0.56)	37 (31)	125 (43)	378 (211)	26 (6)
Zone 3A	2.57 (0.53)	-3.76 (1.16)	49 (38)	100 (74)	374 (277)	29 (9)
Zone 3B	2.31 (0.23)	-4.43 (0.71)	19 (18)	nm	nm	nm

Values in bracket represent one standard deviation; nm indicates that no samples were measured.

FIG. 11.—Representative correlation between the percentage of dolomite vs $\delta^{18}\text{O}$ in 15 consecutive samples in zones 1, 2A, 2B, and 3 in the bulk samples from Well C. In Zone 1 there is a positive correlation, as would normally be expected from mixtures of dolomite and calcite. In Zone 2A this correlation changes to an inverse relationship, indicating mixture between calcite and an isotopically negative dolomite. In Zone 2B, which contains low concentrations of calcite, there is no statistically significant correlation, while in Zone 3 the correlation changes back to an inverse relationship. These correlations are shown in greater detail in Figures 7, 8, and 9.

within Zone 2B, and an inverse correlation in Zone 3. This changing correlation is represented in Figures 7 and 8 as a windowed regression coefficient. Further discussion of the significance of this change is discussed later in this paper. The dolomite component in some of these samples is therefore isotopically negative relative to the LMC component (Fig. 10C). On the basis of the inverse correlation between the percent dolomite and $\delta^{18}\text{O}$ of the bulk samples, the $\delta^{18}\text{O}$ of pure dolomite in these samples would be approximately -7‰ , similar to the $\delta^{18}\text{O}$ measured in baroque dolomite from other areas in the field. However, despite the similarity between the $\delta^{18}\text{O}$ of the baroque dolomite and the dolomites measured in this study, they may not be related. This conclusion is based on the dissimilarity of the petrography of the baroque dolomites and the absence of any high concentrations in Fe that characterized the baroque dolomites studied by Cantrell et al. (2004).

A more detailed picture of how the change in the $\delta^{18}\text{O}$ relates to the position of the super-k zone can be seen in Figure 9, which shows a 50' (15.2 m) interval from Well B. The rocks in this portion of the core are composed mainly of dolomite, although there are still intervals that are only partially dolomitized. In the upper portion the rock is composed of approximately 50% dolomite and the bulk rock possesses a $\delta^{18}\text{O}$ of approximately -5‰ . In this upper interval there is an inverse correlation between the percentage dolomite and $\delta^{18}\text{O}$, implying mixing with an isotopically negative dolomite. Within Zone 2B, the $\delta^{18}\text{O}$ of the rock increases and the correlation between $\delta^{18}\text{O}$ and the percentage dolomite becomes positive, similar to that expected as a result of dolomite and calcite formed from similar fluids at similar temperatures. Throughout this interval (6560 to 6600'; 1999.5 to 2011 m) the $\delta^{18}\text{O}$ of the dolomite remains relatively positive (-1‰), with occasional excursions to more negative values, usually coinciding with increases in the concentration of calcite. The super-k zone in Well B is an interval between 6581 and 6585' (2005.9 to 2007.1 m), that despite having no measurable calcite shows more negative $\delta^{18}\text{O}$ values. The $\delta^{13}\text{C}$ values in this interval also show slightly more negative values, but there is no perceptible change in the concentration of Sr, Mn, or Fe (see Fig. 13). The absence of calcite within this interval causes the regression coefficient between the concentration of dolomite and $\delta^{18}\text{O}$ to fall to zero. Below the super-k, the $\delta^{18}\text{O}$ of the dolomite becomes more positive, similar to values higher in Zone 2B. The small change in the $\delta^{18}\text{O}$ in the super-k interval suggests the presence of a dolomite formed

FIG. 12.—The $\delta^{13}\text{C}$ vs. $\delta^{18}\text{O}$ of the chemically isolated dolomites measured in this study compared to Pleistocene dolomites from San Salvador in the Bahamas (Supko 1977; Dawans and Swart 1988), Pliocene and Miocene dolomites from hole Clino on the Great Bahama Bank (Swart and Melim 2000), and from modern dolomites in Abu Dhabi (McKenzie 1981). Dolomites from the Arab-D have ranges of $\delta^{13}\text{C}$ values similar to these more modern examples, but a more negative and wider range of $\delta^{18}\text{O}$ values.

from different fluids or at a higher temperature. In Well C, the regression coefficient between the percentage dolomite and the $\delta^{18}\text{O}$ of the bulk rock also decreases to zero within the super-k zone.

Carbon.—The $\delta^{13}\text{C}$ of the bulk rocks ranges between approximately $+1.5$ and $+2.5\text{‰}$. These values are only slightly more negative than that exhibited in modern carbonate platform sediments (Lowenstam and Epstein 1957; Milliman 1974). There is a statistically significant correlation between the percentage of dolomite and the $\delta^{13}\text{C}$ of the bulk rock ($r = 0.64$, $n > 500$, $p \ll 0.01$), a correlation that can be accounted for as a result of the approximate 1‰ enrichment in the $\delta^{13}\text{C}$ of dolomite relative to co-genetic calcite (Sheppard and Schwarz 1970). Under normal circumstances, diagenesis under the influence of meteoric fluids (Matthews and Allan 1978; Allan and Matthews 1982) results in the formation of carbonates with highly negative $\delta^{13}\text{C}$ values, hence the relatively high $\delta^{13}\text{C}$ values of all the dolomites suggest that the rocks experienced alteration in a relatively closed system and were probably not significantly influenced by meteoric fluids prior to dolomitization. Although the variations in $\delta^{13}\text{C}$ are minor, the variations that do exist seem to covary with the correlation between the $\delta^{18}\text{O}$ of the bulk rock and the percentage dolomite. Hence, when there is a positive correlation between the $\delta^{18}\text{O}$ of the bulk rock and the percentage of dolomite, then the $\delta^{13}\text{C}$ of the bulk rock is more positive than when there is a negative correlation between the $\delta^{18}\text{O}$ of the bulk rock and the percentage of dolomite (Figs. 7, 8). This further supports the notion that late-stage fluids, perhaps elevated in temperature and depleted in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$, altered these rocks. Within the super-k zone of Well B, the $\delta^{13}\text{C}$ values of the bulk rock show slightly lower values.

Strontium.—The concentration of Sr was analyzed in the dolomite separates in order to help constrain the nature of the fluids responsible for the dolomitization (Land 1980; Machel and Mountjoy 1986; Budd 1997). Based on previous work (Vahrenkamp and Swart 1990, 1994; Banner 1995; Malone et al. 1996) it is known that stoichiometric dolomites, when formed from fluids with Sr/Ca ratios close to seawater, have a Sr concentration of approximately 100 ppm. Dolomites formed from solutions with more

FIG. 13.—Concentration of A) Sr, B) Fe, and C) Mn in Well B and D) Sr, E) Fe, and F) Mn in Well C. Data are from the chemically isolated dolomites only.

evolved Sr/Ca ratios can have Sr concentrations of 2000 ppm and higher (Land 1973; Humphrey 1988; Swart and Melim 2000). Although it has been suggested that higher-salinity fluids produce dolomites with higher concentrations of Sr (Lucia and Major 1994) and that mixing-zone dolomites can be distinguished upon the basis of a lower Sr concentration (Major 1984; Fouke 1993), the salinity of the fluids *per se* do not have any influence upon the Sr concentration of the dolomite, unless the fluid has evolved to the point at which the Sr/Ca ratio has been altered. The Sr/Ca ratio of a fluid can be increased as a result of recrystallization of carbonate minerals, gypsum, or anhydrite in a closed system inasmuch as all these minerals have distribution coefficients for Sr of less than unity (Veizer 1983; Butler 1973). The Sr/Ca ratio of the fluids can decrease through the precipitation of Sr bearing minerals such as celestite. In addition, lower Sr/Ca ratios might be produced through mixing of calcium-rich freshwater with seawater.

The results from the analyses of the cores measured in this study are shown in Figure 13. These data show that the dolomites possess Sr concentrations close to that expected as a result of formation from an open and rapidly circulating seawater system (Vahrenkamp and Swart 1994). A comparison of the Sr concentration in the dolomites between Zone 2 and 3 is shown in Tables 1 and 2. The difference between Zone 2A and the other areas of the wells is significant at the 99.9% confidence limits. This may indicate dolomitization from more evolved fluids such as those involved in the precipitation of evaporite minerals.

Iron and Manganese.—Iron and manganese have distribution coefficients greater than unity for incorporation into dolomite (Veizer 1983) and therefore relatively high concentrations in the dolomites. Equilibrium concentrations of Mn and Fe in dolomites formed from seawater are approximately 1 ppm and 3–50 ppm respectively. Values reported in the literature for dolomites (Budd 1997; Vahrenkamp and Swart 1994) interpreted as having been formed from seawater are generally higher, suggesting contributions from subaerial exposure surfaces known to have higher concentration

of these elements (Rossinsky et al. 1986) or from the formation of Mn and Fe crusts on the seafloor. Concentrations of Mn measured in this study were similar to those reported by Vahrenkamp and Swart (1994) in dolomites from Little Bahama Bank, which were interpreted as being formed from normal seawater. In contrast, the concentrations of Fe were significantly higher than those reported by Vahrenkamp (1988).

Although the higher values may simply reflect some greater degree of contamination experienced by the rocks during drilling, it is also possible that the higher concentrations indicate a small contribution from higher-temperature dolomites (baroque dolomites) found elsewhere in the Ghawar field that are known to have high concentrations of Fe (Cantrell et al. 2004). However, while concentrations of Mn and Fe were correlated with each other in both wells, they showed no relationship with $\delta^{18}\text{O}$. This absence of correlation with the isotopically negative dolomites seems to suggest that these are not the same as the baroque dolomites that were discovered to have unusually high concentrations of iron ($> 10,000$ ppm) and $\delta^{18}\text{O}$ values of around -7‰ and that there may be more than one type of isotopically negative dolomite.

Geochemical Evidence of Fracturing

The development of the high-permeability zones (super-k) within the Arab-D is of obvious interest in that such zones exert a major control on the oil production within the Ghawar field. It has been recognized from previous work that the super-k is often associated with dolomites in Zone 2A or 2B of the Arab-D (Mitchell et al. 1988; Meyer et al. 2000), although there are rare instances of its association with the top of Zone 3A or to nondolomitized sections. In this study we were able to examine the geochemistry of two wells (wells B and C) in which the super-k occurred in a dolomitized interval (the super-k from well A was not recovered). Dolomites in the super-k intervals appear to be petrographically similar to those above and below, with the exception that there is a greater porosity

and permeability associated with the super-k. The super-k also has a tendency to occur in intervals dominated by the stromatoporoid *Cladocoropsis*. Although the original mineralogy of this species is not certain, and aragonite, low-Mg calcite, and high-Mg calcite have been proposed (Galloway 1957; Stearns and Mah 1987; Yoo and Lee 1993), the specimens present in the studies cores are in most instances leached, leaving molds and cavernous porosity. If the *Cladocoropsis* were originally low-Mg calcite, we might speculate that they would have resisted dissolution and early dolomitization, while if they were aragonite or high-Mg calcite they might have been dissolved very early in the paragenetic sequence. In terms of the stable-isotope geochemistry of the super-k zones, in the two examples that we examined in this study, the zone occurred in intervals with overall more positive $\delta^{18}\text{O}$ values. This enrichment is attributed to the presence of early dolomites that formed from seawater with normal or slightly elevated salinity and at temperatures between 35 and 40°C. Within these zones there is a positive correlation between the $\delta^{18}\text{O}$ of the bulk rock and the percentage of dolomite. In contrast, above and below the zone of high $\delta^{18}\text{O}$ values there is an inverse relationship between the $\delta^{18}\text{O}$ of the bulk rocks and the percentage of dolomite. This implies that the entire sequence was subjected to a later dolomitization event at higher temperature and/or fluids with differing isotopic composition. The zones with the positive $\delta^{18}\text{O}$ values, which may have been partly dolomitized early in their diagenetic history, were already mineralogically stabilized and were less susceptible to later recrystallization. In order to examine the association between the percentage of dolomite and $\delta^{18}\text{O}$ values in the super-k zones in more detail, we carried out a moving correlation analysis between these two variables ($\delta^{18}\text{O}$ in the bulk rock and the percentage dolomite). In this technique a regression coefficient is calculated in a window of a specific number of samples (10). The initial sample within the window is then incremented by one and a second regression coefficient calculated. In this manner the spatial variation in the regression coefficient can be determined (the number of degrees of freedom in this calculation is 9 and hence the regression coefficient that is statistically significant at the 95% confidence limit is 0.602). If the relationship between the two variables being correlated is the same throughout the sequence, the correlation coefficient will not vary significantly. However, a changing correlation coefficient indicates differing relationships between the percentage dolomite and the $\delta^{18}\text{O}$ in different portions of the core. The calculated regression coefficients are shown in Figures 7 and 8 and indicate that in the center of the zone that has the higher $\delta^{18}\text{O}$ values and in which the super-k zone occurs, the positive relationship between the $\delta^{18}\text{O}$ of the bulk rock and the percentage of dolomite disappears; the regression coefficient changes from a value of approximately +0.7 to near zero. In Well B this change in the correlation coefficient reflects the fact that the rocks surrounding the super-k zone are composed of 100% dolomite, but trend from a $\delta^{18}\text{O}$ value of -1‰ outside the super-k zone to -3‰ adjacent to the zone (Fig. 11).

A correlation between the percentage of dolomite and $\delta^{18}\text{O}$ in such a situation is obviously meaningless as there is no variation in the concentration of dolomite. In Well C, the correlation coefficient also tends towards zero in the super-k zone, although in the case of Well C the rocks are composed of mixtures of dolomite and calcite. Therefore, the poor regression reflects mixing between calcite and dolomite with both more positive and negative $\delta^{18}\text{O}$ values. Hence there is a suggestion that the super-k intervals, which have overall more positive $\delta^{18}\text{O}$ values, also contain a later-stage higher-temperature-derived dolomite. The presence of quartz within the super-k zone in Well C, which is often associated with hydrothermal solutions (Rimstidt 1997) also supports the notion that there is an influence of high-temperature solutions within the super-k zone. Although high-temperature baroque dolomites have been reported in other cores from the Arab-D (Cantrell et al. 2004), these are associated with significantly higher concentrations of Fe than measured in the dolomites isolated from the super-k samples in this study and therefore the fluids responsible for

FIG. 14.—Temperature vs. $\delta^{18}\text{O}$ (PDB) for various $\delta^{18}\text{O}$ values of fluids (SMOW), assuming that the $\delta^{18}\text{O}$ of dolomite is 3‰ heavier than calcite precipitated under identical conditions (Land 1980, 1983). Box A shows the isotopic composition and estimated water value for the early formed dolomites in Zone 2B. Estimates of the salinity are based on an ice-free world and temperature between 30 and 40°C. Box B shows dolomites with the most isotopically negative values recognized in this study. Estimates of the salinity for the formation fluids are taken from Cantrell et al. (2004), and salinity vs. $\delta^{18}\text{O}$ (SMOW) of fluids are estimated from the known relationships between $\delta^{18}\text{O}$ and salinity in evaporating brines.

the super-k may be different from those which formed the baroque dolomites.

In order to calculate the temperature of the formation of the dolomite using the $\delta^{18}\text{O}$ values, it is necessary to have some knowledge of the $\delta^{18}\text{O}$ of the solutions from which the dolomites formed. Some idea of this value can be obtained from the fluid-inclusion data measured on other dolomite samples from the Arab-D (Cantrell et al. 2004), which suggest salinities of 6 to 7 times that of seawater. Although increases in salinity do not equate directly to increases in the $\delta^{18}\text{O}$ of water (Gonfiantini 1986), it is a good probability that the $\delta^{18}\text{O}$ of the formation fluids was at least $+2\text{‰}$ and maybe as high as $+4\text{‰}$ (SMOW). Utilizing this estimate of the $\delta^{18}\text{O}$ of the water and $\delta^{18}\text{O}$ of the most negative dolomites measured in the super-k zone, then it is possible to calculate a temperature of formation between 80 and 100°C (Fig. 14).

Although the presence of fractures has been previously proposed as a mechanism for creating super-k properties (Meyer et al. 2000) and there was some evidence of fracturing evident in thin sections as mentioned earlier, physical evidence for the presence of fractures is largely absent in the cores themselves. The lack of fractures is not an obstacle to our hypothesis, because normal cores often fail to capture fractures which are mainly oriented vertically.

Implications for the Formation of Porosity and Permeability in Dolomitized Reservoirs

It was originally suggested that dolomitized rocks are generally better reservoirs than limestones because, during certain dolomitization reactions, two moles of calcite when converted to one mole of dolomite, lead to a net increase in porosity (Beaumont 1837; Weyl 1960; Murray 1960). However, a large number of subsequent workers have suggested other reasons which may govern porosity changes during dolomitization. These include the characteristics of the precursor sediment (Lucia and Major 1994), subsequent leaching (Purser et al. 1994), and cementation (Moore 1989; Moore and Heydari 1993). In the situation described in this paper, we propose that early partial dolomitization provided a template which not only led to

preferential fracturing, allowing later diagenetic fluids to penetrate the formation, but also provided a mineralogy that resisted later dissolution by these fluids. Solutions penetrating these fractured units dissolved away remaining nondolomitized material, creating a porous and permeable reservoir, which is now mainly dolomite. It is well established that fracturing affects carbonate reservoirs (Roehl and Choquette 1985), although its influence upon the porosity remains uncertain (Moore 2001). In the case of the Arab-D, it is proposed that fracturing is concentrated in the early-dolomitized units. The enhanced porosity which is observed now in these dolomitized units is not a result of the dolomitization process *per se*, but of the physical and chemical characteristics which dolomitization imparts to the rock, which in turn influences how later solutions impact the rock.

Paragenetic Sequence

On the basis on the petrographic and geochemical data presented above we propose the following paragenetic sequence.

A. Dolomitization associated with Zone 3 of the Arab-D: A portion of the dolomite in Zone 3 is considered to have formed approximately contemporaneously with deposition and is associated with the presence of numerous firmground surfaces throughout the zone. In the example shown in Figure 7, the concentrations of dolomite were higher below seven of the NDS, while they show no increase below a further four. In addition, one interval which showed an increase in dolomite was not apparently associated with any nondepositional surface. The absence of an association between increases of dolomite and surfaces is easily explained using the model of Swart and Melim (2000), which suggests that the amount of dolomite associated with a surface is proportional to the hiatus represented by the surface. Hence NDS that represent only a short duration contain minimal amounts of dolomite. The instance in which an increase in dolomite is not associated with a NDS is difficult to explain using this model, but it is possible that physical evidence of the hiatus was not recognizable in the core. In this regard it should be noted that the core in the interval around 7380' (2249.4 m) in Well C (Fig. 4) is jumbled, and it is entirely probable that a surface may not have been recognized for this reason. Typically, concentrations of dolomite are high immediately below these surfaces and decrease with downward distance from the surfaces, indicating an association with the surfaces or the overlying depositional unit (Fig. 4). The dolomites in Zone 3 are euhedral, non-fabric-preserving, and approximately 100 μm in size. It is suggested that these dolomites formed in association with the NDS that contain abundant burrows. In many cases these burrows are preferentially dolomitized, first as small nonstoichiometric rhombs (10 to 50 μm) that subsequently coarsened to their present size. This is consistent with the observation that in Zone 3 dedolomitization is restricted to the cores of dolomite rhombs that correspond to the earliest, calcian dolomite growth. The mechanism of formation is postulated to be similar to that documented by Swart and Melim (2000) for dolomites forming off the margin of the Great Bahama Bank and proposed for dolomitization in other fine-grained sedimentary sequences such as the Monterey (Baker and Burns 1985; Compton and Siever 1986). In this model dolomitization is promoted by the oxidation of organic material by O_2 and sulfate during very shallow burial, which causes dissolution of the precursor carbonates. The Mg^{2+} ions necessary for dolomitization are provided by diffusion from the overlying seawater, and therefore the concentration of dolomite is highest closest to the NDS and decreases with depth. Originally these dolomites would have had $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ ($\delta^{18}\text{O} +2$ to $+4\%$ and $\delta^{13}\text{C} +2$ to $+4\%$) values reflecting deposition from normal marine fluids during Jurassic times ($\sim -1\%$ SMOW). The fact that the dolomites now have quite negative $\delta^{18}\text{O}$ values (-6 to -6.5%), suggests later alteration by warm fluids during burial or recrystallization in a closed system in a normal geothermal gradient. Because diagenesis would have taken place in a closed system, this process would have reset the oxygen isotopic systematics, but not the carbon. Hence the dolomites still possess $\delta^{13}\text{C}$ values typical of modern

marine sediments. A closed-system dolomitization might also be expected to lead to a higher Sr concentration (Swart and Melim 2000). However, if dolomitization occurred close to the nondepositional surface it might be expected that the Sr concentration in the dolomite would still be close to values typical of formation from normal seawater.

B. Dolomitization in Zone 2: A second phase of dolomitization affected the generally more grain-dominated sediments characteristic of Zone 2B. Although original depositional fabrics are often difficult to discern as the original textures have been masked by dolomitization, these sediments probably represent a slightly shallower-water facies compared to the sediments of Zone 3. The dolomites formed in these sediments are principally non-fabric-preserving in nature, but the crystals are larger in size (200 to 500 μm) compared to those found in Zone 3 and they possess significantly more positive oxygen isotope compositions ($\delta^{18}\text{O} = -2$ to -1%). Some of these dolomites possess extremely high permeability and porosity and are responsible for some of the high-flow zones (super-k) encountered in this field, while others possess low permeability and porosity. We interpret that the dolomites in Zone 2 formed through the reflux of higher-salinity fluids down into these sediments, as evidenced by the presence of dolomites with relatively positive $\delta^{18}\text{O}$ values in this zone; the dolomites in Zone 1, which are intimately associated with evaporites, have $\delta^{18}\text{O}$ values similar to those in Zone 2.

Two possible time periods for this reflux are possible. The first is immediately following the deposition of Zone 2B, when shallow seas could have been conducive to the formation of hypersaline brines, perhaps in a situation analogous to the Bahamas. The second period of possible reflux is during or shortly after the deposition of the evaporites overlying Zone 1. Although there is some evidence to support the first time period of dolomite formation in Zone 2B (specifically the paucity of dolomite in Zone 2A), it is not certain that the fluids present on the platform at that time would have had sufficient thermodynamic drive to produce the amount of dolomite observed. In contrast, fluids originating from the overlying evaporites would have been powerful dolomitizing agents. We propose that in order to deliver the fluids to Zone 2B without influencing Zone 2A, a series of fractures and joints developed in Zone 1 and 2A coincident with deposition of the CD evaporite (at the boundary between Arab D and Arab C members). As a result of non-fabric preserving dolomitization of Zone 2, the sedimentology is not precisely known. We suggest, that within this zone, the grainstone units were more pervasively dolomitized at this time, while the muddier units may have been only partially dolomitized. The *Cladocoropsis* fossils that characterize the super-k zone are now mainly dissolved and present only as molds. If these fossils were originally composed of high-Mg calcite (Yoo and Lee 1993) or aragonite (Stearns and Mah 1987) they may have been dissolved early in the paragenetic sequence, creating the large molds that are characteristic of this zone. Alternatively, if they were originally low-Mg calcite they were probably not dolomitized early but rather dissolved during a later dolomitization event.

C. Dolomitization in Zone 1: Dolomites associated with Zone 1 are petrologically distinctive (fabric preserving) and are intimately associated with the evaporites in this zone. The association of dolomites and evaporites has been well documented in the literature (Kinsman 1964; Bush 1973; Hsu and Schneider 1973). It is suggested that elevation of the Mg/Ca ratio in the pore fluids as a result of the precipitation of aragonite, gypsum, and anhydrite promotes the precipitation of dolomite. Such processes take place in an evaporative setting associated with waters of more positive $\delta^{18}\text{O}$ values (McKenzie 1976; McKenzie et al. 1980). Hence by inference the dolomites in Zone 1 that possess relatively positive $\delta^{18}\text{O}$ values (-1%) are interpreted as being formed by saline to hypersaline fluids at or very shortly after the time of deposition.

D. Later Stage Dolomitization: During increasing burial depth the sediments became compacted, with the formation of stylolites in the muddier portions of the section (Zone 3), and were fractured in the portions of the section that were dolomitized early (Zone 2B). These fractures served as

conduits for alteration by later fluids. The influence of more than one dolomitization event is evidenced by the high variability of dolomite crystal size within a sample, as well as the range in $\delta^{18}\text{O}$ values. These late fluids were elevated in temperature and moved preferentially through the fracture system (which probably developed fairly early, at or shortly after the time of deposition of the evaporite) and dissolved a portion of the sediment that was not dolomitized. Hence within Zone 2B, parts of the sediments that were only partially dolomitized were leached and the matrix removed leaving behind a sucrosic dolomite possessing very high porosity and permeability. Fragments of *Cladocoropsis* could have been leached at this time leaving molds that are typical of this zone. These zones, which eventually formed the super-k intervals, occurred preferentially in the muddier portions of the sediments. In contrast, the more grain-dominated sediments may have been completely dolomitized early and were therefore less affected by this late-stage alteration event. Evidence for the association between the super-k zones and the late-stage dolomitization event is evident in the $\delta^{18}\text{O}$ composition of the dolomites associated with the super-k zones in wells B and C. This zone always shows an enrichment in the $\delta^{18}\text{O}$ values, but it is surrounded by dolomites depleted in $\delta^{18}\text{O}$. This depletion is consistent with alteration by late-stage hydrothermally derived fluids. It is also possible that these late-stage fluids also altered the dolomites in Zone 3 that have $\delta^{18}\text{O}$ values as negative as -6.5% . Late-stage alteration is also suggested by the presence of dedolomite in Well A. Fluids undersaturated with respect to both calcite and dolomite could have preferentially leached less stable portions of the dolomites, (Fig. 5F) and replaced the dolomite in some instance with LMC.

An alternative explanation is that the dolomites with the more negative $\delta^{18}\text{O}$ values simply recrystallized or formed in the presence of a geothermal gradient associated with increased burial. Assuming an original oxygen isotope composition for these dolomites of approximately $+3\%$, the present $\delta^{18}\text{O}$ suggests an alteration temperature of about 50 to 60°C . These late-stage alteration features are conspicuous in that they differ strongly between the individual cores. This supports the notion of fractures being the conduits for the late fluids, even though fractures have not been extensively observed in these cores.

SUMMARY

- At least four episodes of dolomitization have been identified influencing the Arab-D in the Ghawar field. The earliest episodes occurred contemporaneous with deposition in Zone 3, 2, and 1. Dolomitization at this time was probably not complete. In Zone 3, dolomitization was associated with periods of nondeposition and the Mg^{2+} was supplied by diffusion from the overlying seawater. The amount of dolomite formed was linked to the period of nondeposition. Within Zone 2 the fluids responsible for dolomitization could have been derived from slightly evaporated seawater or fluids modified by the formation of evaporites. The fluids were driven through the sediments by gravity-driven flow. In Zone 1 dolomitization occurred contemporaneously with the deposition of the evaporites.
- After deposition, early dolomitization, and further burial, compaction of the sedimentary column occurred. The early dolomitized portions were less influenced by compaction, but were susceptible to early fracturing.
- The second episode of dolomitization occurred from evolved fluids circulating through the platform. This episode affected all the sediments and partially leached and further dolomitized the sediment that was partially dolomitized earlier. These fluids were slightly elevated in temperature compared to the fluids responsible for the early dolomitization.
- During later burial diagenesis, dolomites formed as a result of recrystallization in the presence of the normal geothermal gradient.
- During burial and tectonic movements, major fracturing of the reservoir occurred. This fracturing preferentially affected cemented units, such as the early formed dolomites. Sediments that were not dolomitized early

were subjected to formation of stylolites and loss of porosity. Fluids passing along these fractures influenced the early-dolomitized units, leaching remaining limestones and imparting a subtle geochemical signature to the remaining rocks.

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