PALAEOKARSTS AND PALAEOKARSTIC RESERVOIRS

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FOREWORD

This course book aims at presenting a review of palaeokarsts specifically oriented for the exploration geologist.

The first chapter reviews the dissolution processes, hydrology and porosity evolution of karst systems. Karst is one of the few diagenetic systems in which a strong hydrogeological framework exists, allowing some predictive component for porosity distribution.

Chapter 2 introduces palaeokarsts, covering such topics as terminology, recognition, controls and associations.

Chapter 3 relates to reservoir aspects of palaeokarsts while Chapter 4 provides some case studies of hydrocarbon-bearing palaeokarstic reservoirs.

The use of the term karst in this book needs some qualification. Carbonate rocks are prone to dissolution and a critical distinction must be made between general dissolution and karstic processes. The term karst is used here to describe solutional porosity development, where the pore/conduit diameter exceeds 5-10mm allowing flow to change from laminar to turbulent, creating more rapid dissolution. Such dissolution is not limited to purely meteoric groundwaters.

A detailed review of karst processes and geomorphology has been provided by Ford and Williams (1989). Shorter reviews on karst systems and hydrochemistry, written for geologists, have been given by Ford (1988) and Lohmann (1988). Choquette and James (1988) provide a general introduction to karst and palaeokarst and they have also given a very useful review of meteoric diagenesis in carbonates (James and Choquette, 1984).

Two important volumes on the topic have appeared. James and Choquette (1988) have compiled a series of reviews and case studies on karstic processes and palaeokarsts, which is essential reading. Bosak et al. (1989) have provided a global review of palaeokarst occurrence.

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CHAPTER 1

KARST PROCESSES, HYDROLOGY AND POROSITY EVOLUTION

by P.L. Smart & F.F. Whitaker

Karst is a term used to describe distinctive terrains whose landforms and hydrology result from a combination of high rock solubility and well developed secondary porosity (Ford & Williams, 1989). From a diagenetic point of view, karstification can be considered as one of the range of processes affecting sediments, especially carbonates and evaporites. It constitutes a distinctive 'diagenetic facies' (Esteban & Klappa, 1983). The following sections will stress the wide range of processes and settings in which carbonate dissolution and karst features can occur.

This chapter will examine the processes responsible for karst development, the 'realms' (process domains) of karst formation and their hydrology and, lastly, the evolution of porosity during karstification.

1.1 CARBONATE DISSOLUTION PROCESSES

An understanding of the processes of carbonate dissolution is essential in predicting the extent and distribution of secondary porosity in palaeokarst terrains. The mechanisms driving dissolution of carbonates have been divided into two categories by James and Choquette (1984).

1) Water Controlled - involving equilibrium between water and carbonate minerals, for instance calcite dissolution by carbonic acid.

2) Mineral Controlled - involving the conversion of a thermodynamically unstable mineral to one that is stable, for instance the transformation of biogenic high magnesium calcite to low magnesium calcite.

Five water controlled reactions may be of significance in palaeokarst terrains:

1. dissolution by carbonic acid.

2. dissolution as a result of oxidation of organic matter.

3. dissolution by mixing of waters of different chemistry.

4. dissolution resulting from fluid temperature changes.

5. dissolution by gases or fluids expelled from adjacent basins.

Each of these, together with the mineral controlled reactions, gives rise to a characteristic porosity profile with depth below the surface.

1.1.1 Dissolution by Carbonic Acid

Process In many environments carbonic acid is responsible for the majority of carbonate dissolution. The process is summarised in reaction (1), but it is instructive to consider the individual steps involved. Calcium carbonate is sparingly soluble in pure water.

\[
\text{CaCO}_3 + \text{CO}_2 + \text{H}_2\text{O} \rightarrow \text{Ca}^{2+} + 2\text{HCO}_3^- \quad (1)
\]

At 10°C only 6 mg/l of calcite will dissolve - a solubility lower than that of quartz. The solubility of aragonite is significantly higher, while that of dolomite is substantially lower. The carbonate ions formed react with hydrogen ions from the dissociation of water, raising the pH.

\[
\text{H}_2\text{O} = \text{H}^+ + \text{OH}^- \quad (3)
\]

\[
\text{CO}_3^{2-} + \text{H}^+ = \text{HCO}_3^- \quad (4)
\]

Thus any acidity contributing hydrogen ions will drive equation (4) to the right, disturbing the equilibrium of equation (2), and causing more limestone to dissolve. For instance, sulphuric acid derived from the oxidation of iron pyrite in shales overlying the Carboniferous Limestone.
in South Wales is important in generating a well developed interstratal karst (see Chapter 2).

Carbonic acid is formed by dissolution of gaseous carbon dioxide.

\[ \text{CO}_2(\text{g}) = \text{CO}_2(\text{aq}) \]  
\[ \text{CO}_2(\text{aq}) + \text{H}_2\text{O} = \text{H}_2\text{CO}_3 \]  
\[ \text{H}_2\text{CO}_3 = \text{H}^+ + \text{HCO}_3^- \]  

Carbon dioxide is ubiquitous, and even the low amount present in the atmosphere (0.03%) is sufficient to dissolve about 70 mg/l of calcite at 10°C. The amount of carbonic acid dissolved is dependent directly on the partial pressure of carbon dioxide (PCO2) in the gas phase, and inversely upon temperature (Fig. 1.1). However, the temperature effect is small compared with the range of PCO2 expected in carbonate terrains (see Thrailkill, 1976, for further details of the dissolution process).

**Soil Carbon Dioxide.** Most soil atmospheres have a PCO2 in excess of the atmospheric value because of in situ production of CO2 by root respiration, and microbial decomposition of organic matter.

\[ \text{C}_6\text{H}_{12}\text{O}_6 + 6\text{O}_2 = 6\text{CO}_2 + 6\text{H}_2\text{O} \]  

Concentrations as high as 0.1 atmospheres have been reported, but typical values are in the range 0.001 to 0.025 atmospheres (Smith & Atkinson, 1976; Miotke, 1974). The carbon dioxide concentration is dependent on the balance between production and removal of carbon dioxide by upward diffusion to the soil surface (De Jong & Schappert, 1972). Thus thick soils with low gas permeability have the highest carbon dioxide concentrations in a given area, as shown by Crowther (1986) for karst soils in Peninsular Malaya. This is important in causing spatial differences in the rate of bedrock dissolution, and gives rise to topographic forms such as tower karst. In this terrain, rates of solution on the limestone hills, which have thin soils, are much less than on the intervening plains, which have much thicker (and often wetter) soils (155 compared to 85mm/ka, Crowther, 1986).

Carbon dioxide production varies with plant activity and the rate of bacterial decomposition of organic matter. Both are controlled predominantly by temperature and soil moisture, and in many climates there is a strong seasonality in carbon dioxide concentrations (De Jong & Schappert, 1972; Miotke, 1974). Regional variations in temperature also affect carbon dioxide production, and Harmon et al. (1975) showed for North America that the average annual PCO2 of groundwaters was directly related to mean annual temperature (Fig. 1.2).

Whilst laboratory experiments suggest temperature is over twice as important as soil moisture in controlling soil carbon dioxide concentrations, it is difficult to separate these two factors in field data because of co-variation. Certainly in some areas such as Madagascar, soil moisture provides the major seasonal control. Brook et al. (1983) thus used actual evapotranspiration (AET), which combines available energy (temperature) and moisture as a predictor for a world model of growing season soil CO2 concentrations. It is notable that many of the world’s most karstified areas such as South East Asia, Mexico and the Caribbean fall within the areas with the highest predicted soil PCO2.

**Rates of Dissolution.** The global variation of carbonate dissolution is not however wholly dependent on the chemical potential for dissolution, but also depends on the effective runoff (precipitation minus evapotranspiration), which controls the amount of water passing through the system. Measurements of denudation rates in limestone areas show a linear dependence on runoff, with bare limestone areas (arctic and alpine areas) having less erosion than soil covered terrains in the tropics and temperate areas (Fig. 1.3, Smith & Atkinson, 1976). High runoff is generally associated either with mountainous areas of considerable relief (A in Fig. 1.3) or with the seasonally humid tropics (B in Fig. 1.3).

White (1984) has modelled the effects of runoff, soil PCO2 variation and
Fig. 1.1 Equilibrium concentration of calcium carbonate in water for different PCO$_2$ and temperature.

Fig. 1.2 Variation of mean PCO$_2$ of spring waters with temperature for North American karst terrains (modified from Harmon et al. 1975).
temperature on disso1utional erosion rates (Fig. 1.4). The direct effect of temperature is relatively minor (c. 30% decrease with a change from 5° to 25°C). Denudation rate varies as the cube root of soil PCO₂, thus, for the two orders of magnitude range predicted by the Brook et al. model, global denudation rates may vary by a factor of five. In contrast, because denudation varies directly with runoff, a variation of the latter from 300 - 3000 mm/a gives an order of magnitude change in erosion rate. This model thus demonstrates the dominant role of runoff in controlling denudation rates (and thus porosity generation).

The somewhat higher denudation rates actually measured for bare limestone terrains, compared with those predicted by White (1984), may be due to other sources of acidity, for instance organic acids produced by fungi on the bare rock surfaces. A second possibility is that bacterial oxidation of organic matter transported from the surface into the vadose zone may occur (Atkinson, 1977; see also section 1.1.2). Measurements in boreholes suggest an increase in CO₂ with depth. This pattern could however be explained by downward diffusion from the soil during the growing season, followed by upward movement during the dormant season when the diffusion gradient is reversed. These processes are important in permitting continued bedrock dissolution in the vadose zone.

Dissolution Kinetics. The kinetics of calcite dissolution are complex and depend on the degree of undersaturation and the PCO₂ (Berner & Morse, 1974). As equilibrium is approached, the rate of dissolution falls by 3 to 4 orders of magnitude. Thus field observations show rapid dissolution at the soil-bedrock interface and in the top 10m of the bedrock, but some dissolution also continues at depth (Fig. 1.5). Observations in many maturely karstified fissured karst areas confirm this pattern, with 10 to 20% of the dissolution in the sub-soil zone, 50 to 60% in the uppermost bedrock, and 20 to 40% at depth (Smith & Atkinson, 1976). In intergranular flow aquifers, equilibrium would be approached at much shallower depths due to the greater rock/water contact and slower transmission rate, but detailed studies are lacking.

This pattern of dissolution gives rise to a distinctive pattern of porosity with depth, which is particularly evident in dense, well-cemented limestones which have been re-exposed after burial and therefore develop unloading fractures parallel to the surface (see section 1.3.2). The top 5m of the unsaturated zone (the subcutaneous zone) has a high density of solution openings, the frequency of which reduces with depth (Fig. 1.6). Drainage from this zone is predominantly via solutionally enlarged fissures, which pass vertically down through the vadose zone. These form a focus for shallow drainage in the subcutaneous zone, the radial concentration of flow giving rise to increased solution adjacent to the fissures, and the development of closed depressions (Williams, 1985). In massive maturely karstified limestones, the recharge is predominantly via these shaft routes, and there is relatively little other dissolutional development in the vadose zone resulting in a marked heterogeneity. Where there is a high density of fractures, perching and lateral movement in the subcutaneous zone is reduced, and a more uniform solutional porosity develops within the vadose zone. This pattern is more characteristic of newly stabilised limestones with intergranular flow.

Fossil examples of this style of porosity development are reported widely below unconformity surfaces developed on stabilised carbonates. Features commonly include a clay-rich regolith, often with a concentration of the acid insoluble material from the limestones (Meyers, 1988). This material may infill solutionally enlarged joints and grykes, which reduce in both width and frequency below the unconformity (Kahle, 1988).

Summary

1. Carbonic acid generated within the soil is the major agent driving dissolution in many modern karst terrains.

2. Soil thickness, mean annual temperature, effective rainfall and
Fig. 1.3 Denudation rates for soil covered and bare limestone terrains. A = Julian Alps, Yugoslavia, B = Gunung Mulu, Sarawak (modified from Smith & Atkinson, 1976). Note change in observed relationship for soil covered terrains if solution rate is corrected for presence of non-carbonates in the catchment.

Fig. 1.4 Effects of temperature, PCO$_2$ and effective precipitation on denudation rate (modified from White, 1984).
Fig. 1.5 Variation of mean annual calcium hardness (and range) with depth below ground for waters sampled in a Mendip Cave. Note, substantial dissolution occurs at limestone/soil interface but there is also progressive dissolution with depth.
seasonality control carbon dioxide concentration in the soil.

3. Total annual dissolution is controlled more by effective rainfall (runoff) than by soil PCO₂.

4. Dissolution is concentrated within the uppermost part of the bedrock giving a marked reduction of porosity with depth.

1.1.2 Oxidation of Organic Matter

Process Organic matter may be transported into carbonate rocks in particulate or dissolved states, but concentrations of the latter are generally low (less than 2 mg/l DOC). Particulate material is rapidly filtered out during intergranular flow, thus it is only in fractured and maturely karstified carbonates that this process becomes important. However, in young carbonates significant organic material may be incorporated into the sediment at or subsequent to deposition, either as detrital fragments or as organic sheaths in biogenic carbonates.

Organic matter may be decomposed by a sequence of bacterially mediated reactions. Those giving the highest energy yield being driven to exhaustion before oxidation switches to the next reaction in the sequence (Coleman, 1985; Fig 1.7; Table 1.1).

Aerobic oxidation is dominant where the rate of oxygen consumption may be balanced by either diffusion or advection of oxygen into the system. This is the major process occurring in unconfined carbonate aquifers, and in the vadose zone, and produces carbonic acid which drives carbonate dissolution.

$$C_6H_{12}O_6 + 6O_2 = 6HCO_3^- + 6H^+ \quad (9)$$

In many carbonate aquifers the sub-oxic reactions are of only minor importance because concentrations of nitrate are low, and manganese and iron are present in limited amounts. Thus in many situations, especially in sediments, organic degradation passes rapidly from oxic to anoxic conditions, with sulphate reduction becoming dominant (Edmunds & Walton, 1983).

$$C_6H_{12}O_6 + 3SO_4^{2-} = 6HCO_3^- + 3H_2S$$
$$= 3H^+ + 3HS^- \quad (10)$$

Dissociation of H₂S occurs above pH 7 to yield hydrogen ions, but H₂S is also able to diffuse from the site of production through the solution along a concentration gradient established by reaction with an oxidising agent. This may generate further acidity to drive carbonate dissolution.

$$H_2S + 2O_2 = 2H^+ + SO_4^{2-} \quad (11)$$

Because sulphate concentration is high in marine derived waters, sulphate reduction may generate significant acidity, whereas in fresh waters, exhaustion of sulphate leads to methanogenesis.

$$C_6H_{12}O_6 + 3H_2O = 3CH_4 + 3HCO_3^- + 3H^+ \quad (12)$$

This fermentation reaction is only observed in the most distal parts of confined aquifers, or in deep basinal groundwaters. In deep groundwaters, high temperatures (above 80°C) also permit direct non-microbial sulphate reduction.

Where reduced waters discharge into the ocean, as for instance along the base of the Florida escarpment (Paul et al., 1984), chemosynthesis of organic matter may generate further acidity.

$$CO_2 + H_2S + O_2 + H_2O$$
$$= CH_2O + 2H^+ + SO_4^{2-} \quad (13)$$

This may be of significance in the development of the erosional platform margins typical of the present Bahamas. Decarboxylation is a thermally driven process which can only occur at elevated temperatures and is normally associated with deep burial. It will be considered further in section 1.1.5.

Because organic material washed from the surface may be suspended in the density gradients of the transition zone from fresh to saline waters in an isolated carbonate platform, aerobic oxidation and sulphate reduction are important in causing carbonate dissolution in this zone.
Fig. 1.6  A) Variation of permeability with depth in the unsaturated zone at Corconne, Languedoc, France, from 0 to 8 m. B) Caliper log of borehole in unsaturated zone, Eastern Mendip Hills, England, showing test sections isolated by packers and their hydraulic conductivity.
<table>
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<tr>
<th>Environment</th>
<th>Reaction</th>
<th>Oxidant</th>
<th>Products&lt;sup&gt;1&lt;/sup&gt;</th>
<th>Energy&lt;sup&gt;2&lt;/sup&gt; Yield (KJ)</th>
</tr>
</thead>
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<td>Oxic</td>
<td>Aerobic oxidation</td>
<td>O₂</td>
<td>H⁺</td>
<td>530</td>
</tr>
<tr>
<td>Sub-oxic</td>
<td>Manganese reduction</td>
<td>2MnO₂</td>
<td>2MnO 3OH⁻</td>
<td>485-515</td>
</tr>
<tr>
<td></td>
<td>Nitrate reduction</td>
<td>NO₃⁻</td>
<td>N₂/NH₃ OH⁻</td>
<td>460-505</td>
</tr>
<tr>
<td></td>
<td>Ferric iron reduction</td>
<td>2Fe₂O₃</td>
<td>4Fe²⁺ 7OH⁻</td>
<td>220-235</td>
</tr>
<tr>
<td>Anoxic</td>
<td>Sulphate reduction</td>
<td>1/2SO₄²⁻</td>
<td>1/2S²⁻ H⁺</td>
<td>65</td>
</tr>
<tr>
<td></td>
<td>Methanogenesis</td>
<td>-</td>
<td>1/2CH₄ 1/2H⁺</td>
<td>60</td>
</tr>
<tr>
<td></td>
<td>Decarboxylation</td>
<td>-</td>
<td>R-CH H⁺</td>
<td>60</td>
</tr>
</tbody>
</table>

**Notes:**
1. In addition to HCO₃⁻ and H₂O
2. Per mol organic carbon

**Table 1.1** Zonal reaction scheme for organic matter degradation.
Fig. 1.7  Redox and dissolved species present in groundwater with distance down flow-path for the saturated zone in the Lincolnshire Limestone (fissure flow) and Berkshire Chalk (intergranular and fracture flow) (modified from Edmunds et al. 1974).

Fig. 1.8  Decrease in hydraulic conductivity with formation depth for carbonates in Puerto Rico (see text).
In many carbonate platforms wind-blown dust is an important component of the soil, and may provide a source of iron when translocated via fissures into the saturated zone. This is of importance because, where iron is present, hydroxyl ions are produced on sulphate reduction (Coleman, 1985), and precipitation of iron sulphide (Smart et al., 1988a), and carbonate cement may occur (Pigott & Land, 1986).

Controls

In karstified aquifers with fissure and conduit flow and significant vegetation growth on the surface, transport of organic matter to depth provides a significant mechanism for driving dissolution deep within the saturated zone of the aquifer (Plummer, 1977; Edmunds & Walton, 1983). It is also important in young carbonate sediments, such as those studied by Morse et al. (1985). Two climatic controls are important, high temperatures give more rapid degradation and thus greater acid production; and higher effective runoff increases recharge and the net transport of organic matter into the aquifer. These two factors also combine to give high vegetation production, and it is therefore again wet humid tropical areas that have high dissolution rates from this process, as is indicated by the high PCO2 atmospheres in tropical caves.

Exhaustion of transported organic matter and/or oxidants will cause reduced dissolution both with depth, and along the flow path. Dissolutional porosity thus reduces with depth and distance from source (Fig. 1.8), a pattern emphasised by reduction in fracture frequency with overburden depth, and by the more limited groundwater circulation which results.

Summary

1. Aerobic oxidation, sulphate reduction and methanogenesis can cause carbonate dissolution, which is significant at depth in the aquifer, where other dissolution processes are less important.

2. There is a general reduction of porosity with depth below the water table.

3. These processes are most active in young and organic rich carbonates, or in maturely karstified fractured aquifers with high recharge, high temperatures and abundant surface vegetation.

1.1.3 Mixing Corrosion

Process

The potential for dissolution by the mixing of waters of differing composition has been demonstrated theoretically by Runnels (1969) and Plummer (1975). The non-linear relationship between equilibrium calcium concentration and PCO2 is shown in Fig. 1.9. Two solutions (A and B) of different PCO2 and both at equilibrium with respect to calcite, will mix linearly to produce a solution C which is undersaturated and capable of dissolving additional calcium to reach equilibrium at D. Thus in any carbonate aquifer where waters which have evolved in equilibrium with different PCO2 come into contact, there is the potential for dissolution. For instance, rapid flow through the vadose zone will give recharge waters of different PCO2 to saturated phreatic groundwaters, and dissolution will occur on mixing at the water table.

Mixing effects are not limited to waters of different PCO2, but also occur where there are differences in salinity, temperature and ionic composition of solutions. However, dependent on the shape of the function curve, this may cause supersaturation rather than undersaturation. Plummer (1975) calculated the net effect of these parameters on calcite saturation for mixing of fresh groundwater with sea water, as may occur in the transition zone of a coastal carbonate aquifer (Fig. 1.10). For fresh groundwaters from the Yucatan which are supersaturated with respect to calcite, undersaturation is predicted for sea water additions from 5 to 70%, the asymmetry being a product of the higher saturation of the seawater compared to the freshwater end member. Although this effect is less marked for other end members, mixing corrosion has the potential for generation of considerable dissolutional porosity in coastal and island carbonates.
**Fig. 1.9** Mixing corrosion effect. Solid curve shows the solubility of calcium carbonate at 10°C, with respect to the total carbon dioxide in solution. The mixing of two solutions A and B results in an undersaturated solution at C, and evolution by dissolution of calcite to equilibrium at D. (Modified from Bogli, 1965).

**Fig. 1.10** Change in calcite saturation index as a function of mixing sea-water with groundwater from the Yucatan (modified from Plummer, 1975).
**Fig. 1.11** Vertical variation of salinity, $PCO_2$ and aragonite, calcite and dolomite saturation indices in Evelyn Green's Blue Hole, Bahamas (from Smart *et al.*, 1988a).

**Fig. 1.12** Plan, elevation and passage cross-sections for Aquarius Cavern, Grand Bahama; a blue hole in which development of bedding-plane tubes has resulted from mixing corrosion (Smart *et al.*, 1988b).
Back et al. (1979, 1986) thus invoked mixing of fresh and marine groundwaters at the outlet from regional karst groundwater systems along the Yucatan coast for the generation of major embayments and cave systems. Detailed geochemical and wall rock sampling through the modern transition zone between fresh and saline groundwaters in Blue Holes on South Andros Island, Bahamas, was undertaken by Smart et al. (1988a). The results demonstrate undersaturation of the waters in the transition zone with respect to aragonite and calcite (Fig. 1.11), and corresponding pervasive dissolution of the carbonate wall rock at this level. The zone of maximum dissolution is marked by fretting of the rock surface into a 'Swiss cheese' texture, and suggests that mixing zone dissolution is a very rapid process as sea levels only reached the present position some 5,000 years ago. Extensive horizontal cavern systems, often with abrupt passage terminations, have been described from the Bahamas by Smart et al. (1986b), may result from mixing zone dissolution in the transition zone (Fig. 1.12), see also Mylroie (1988).

In addition to simple inorganic mixing, two organically mediated processes are important in enhancing undersaturation of the transition zone in the Bahamas. Bacterial oxidation of organic material suspended in the density gradients of this zone cause elevated PCO$_2$ (see section 1.2.2). In the anoxic conditions at the base of the transition zone sulphate reducing bacteria are also active and evolve hydrogen sulphide which may oxidise in the upper layers generating sulphuric acid.

Inorganic mixing in the transition zone has been used widely to explain the diagenesis of recent and ancient carbonates. Horizons of vuggy and mouldic porosity, in association with cements of characteristic isotopic signature, indicate extensive dissolution due to mixing in the transition zone (Ward & Halley, 1985; Humphrey, 1987). However, most studies emphasise the transition zone as an environment for dolomitisation or cementation, possibly because cements are easier to characterise than dissolution features. Plummer's (1975) calculations indicate that a mixture of 30 to 75% sea water is undersaturated with respect to calcite, yet oversaturated with respect to dolomite and is thus capable of dolomitisation by replacement. This process, known as the Dora or Schizohaline model (Badiozamani, 1973; Tucker & Wright, 1990), produces thin but laterally extensive lens shaped bodies of dolomite, as described by Choquette and Steinen (1980) from the Mississippian of the Illinois Basin. However, despite its widespread adoption, the model carries uncertainties, and studies of modern mixing zones frequently fail to observe dolomitisation (Vernon, 1969; Smart et al., 1988a).

Controls Mixing zone dissolution is therefore hydrogeologically controlled. Where contrasting waters are brought into contact, dissolution may result (Fig. 1.13), the extent of which is dependent both on the geochemistry of the mixing waters, and the groundwater flux. Thus Sandford and Konikow (1989) have demonstrated that dissolution rates are at a maximum in the upper part of the mixing zone (the less saturated end member), and increase greatly towards the coastal margins where groundwater discharge is greatest (see section 1.1.2). The transition zone at the coastal margins and underlying oceanic carbonate islands is therefore particularly important. Salinity contrasts may also occur between saline surface waters formed by evaporation where ground waters reach the surface, and the adjacent freshwater body. Finally, where fracture flow occurs, contrasts in the chemistry of recharge and phreatic lens waters may be significant.

Laterally continuous horizons of enhanced dissolutional porosity developed over a limited depth range therefore characterise mixing zone corrosion. In oceanic islands, two zones may be present - at the water table and in the transition zone; but only the first zone will be present in aquifers lacking a saline interface.

Summary

1. Dissolution may result where waters of different PCO$_2$, temperature, salinity and ionic composition mix.
Fig. 1.13 A) Zones where mixing corrosion may potentially occur. Note contrast between intergranular and fracture porosity, A-at water table is intergranular flow and B is fracture flow aquifer, note the greater depth in the latter case. C is zone of brackish water discharge near coast where freshwater and saltwater mix. D, at depth, transition zone between fresh and saline groundwaters.

B) Contrast in dissolutional porosity for intergranular aquifer with underlying saline water(1) and fractured freshwater aquifer (line of vertical section marked in 1.13A).
2. Mixing of fresh and saline waters in the transition zone at the margins and underlying oceanic carbonate islands generates a laterally continuous zone of high porosity.

3. Organically mediated reactions may also be important in generating dissolution in the transition zone.

4. In fractured aquifers mixing may result in enhanced dissolution at the water table where recharge and phreatic lens groundwaters come into contact.

1.1.4 Thermal effects

Process The solubility of calcite is greater at low temperature than high under open system conditions (Fig. 1.1). This behaviour is termed retrograde solubility, and has been examined for a range of PCO₂ by Wood (1986), who calculated mass transfer coefficients. These define the mass of the mineral which must dissolve or precipitate in order to maintain thermodynamic equilibrium in the aqueous fluid over a unit temperature change. The rates of mass transfer for calcite are greater for elevated PCO₂ and at low temperature (Fig. 1.14a). For PCO₂ in excess of 10^{-2} atmospheres and temperatures below 100°C, rates are in excess of 3ppm/°C. Thus when groundwaters move from zones of high temperature at depth towards the surface, cooling may result in carbonate dissolution and porosity generation. Conversely cementation may occur where waters warm and precipitation of calcite occurs. Wood (1986) argues that thermally driven mass transfer is most important in aquifers in the deep burial environment where convective circulation will normally occur, but Wood and Hewitt (1986) have also demonstrated its significance in forced fluid flow systems (Fig. 1.15).

Controls Garvin and Freeze (1984) have examined the controls on sub-surface temperatures in regional groundwater circulation systems. Limestones have low thermal conductivities, which may give higher gradients than other sedimentary rocks. However, where they have high permeability, enhanced groundwater flows permit heat transport, reducing geothermal gradients in areas of net recharge, and steepening them in the area of discharge (Fig. 1.14b). This effect is most marked in large scale basin circulations, where the combination of deep circulation and high flow velocities may produce high temperatures in discharging groundwaters.

Effects Bakalowicz et al. (1987) suggest that thermal waters rising along the flanks of the Black Hills uplift, South Dakota, are responsible for the formation of Wind and Jewel Caves (the fourth and tenth longest in the world). The caves developed in a limestone unit at the base of a sequence of younger clastic rocks (Fig. 1.16), and are now fossil, although thermal springs are still active some 15km down dip. Evidence for hydrothermal cave development includes the three dimensional rectilinear maze form guided by fractures, the absence of fluvial sediments, and their lack of a genetic relationship to the surface topography. Mineralisation of the cave voids includes deposition of silica, which has prograde solubility, and therefore precipitates as calcite dissolves to give replacement of the wall rock (Wood, 1986). Jakucs (1976) also describes zones of intense 'pulverisation' and brecciation in dolomites adjacent to hydrothermal outlets near Budapest. Müller (1989) has reviewed the hydrothermal palaeokarsts of Hungary.

Geological examples of this style of dissolutional porosity have not been recognized in carbonates. The characteristic porosity pattern expected would be for enhanced development in more permeable units, with a relatively uniform distribution with depth and possibly evidence for replacement of calcite by silica. The zones of enhanced porosity would also be localised in the palaeo-discharge zone of regional groundwater flow systems.

Summary

1. The solubility of calcite is inversely related to temperature, and shows the greatest change at low temperature and high PCO₂.

2. Forced fluid flow systems will tend to warm in the zone of recharge and cool in
Fig. 1.14 A) Thermal mass transfer coefficients for calcite in the open system CaCO₃ - H₂O - CO₂ as a function of temperature and CO₂ pressure (modified from Wood, 1986).

B) Temperature patterns showing the effects of groundwater flow for varying hydraulic conductivities (A, 10 m/a and B, 500 m/a) and basin dimensions (C, 150 km and D, 600 km) (from Garven & Freeze, 1984).
Fig. 1.15 Plan view of structural and diagenetic contours for a sin (x), sin (y) structure. A shows elevation contours, where the + indicates a dome and - indicates a basin. The base map is overlain by the diagenetic contours in B, which shows the pattern for a flow at 30° clockwise from the X axis. Continuous bands or fairways of porosity enhancement occur. (From Wood and Hewett, 1986.)
the zone of discharge, giving the potential for precipitation and dissolution of calcite respectively.

3. Units of high permeability at depth in large scale basin circulation systems (in which high PCO₂ may also be generated) give high geothermal gradients in the discharge zone, and the greatest potential for calcite dissolution.

1.1.5 Expulsion of Gases and Fluids from Basins

Carbon Dioxide Many deep basin waters have high concentrations of carbon dioxide and significant concentrations of short chain carboxylic acids. These are produced by thermocatalytic degradation of kerogen at temperatures of about 100°C (Carothers & Kharaka, 1978), and may dominate alkalinity in some oilfield waters, driving deep-burial dissolution of calcite. Carbon dioxide, and at higher temperatures, methane gas, may be evolved from these reactions, and move to shallower depths, where enhanced dissolution may result. These processes were cited by Druckman and Moore (1985) to explain the late dissolution-enhanced porosity which provides the dominant oil producing zone in the Smackover Formation, Mount Vernon Field. The porosity is localised in narrow zones parallel to the strike, which have permitted migration of the corrosive fluids.

Hydrogen Sulphide This may also move from deep basins into shallower sediments in the gas phase, where it may react with oxygenated waters to form sulphuric acid. Transport of hydrogen sulphide in regional groundwater flows does, however, appear to be more common. Egemeier (1981) reported the development of simple linear cave systems, where thermal waters bearing hydrogen sulphide gas and hydrocarbons discharge from limestones exposed in breached anticlines in the Big Horn Basin, Wyoming. Much of the cave development appeared to be via oxidation of the evolved gas in thin water films on the walls of air-filled caves. Where mixing of oxidised and reducing waters occurs, dissolution caves may develop over a much larger vertical range (Davis, 1980).

Carlsbad Caverns in the Guadalupe Mountains of New Mexico, includes some of the largest cave chambers known, and contains extensive deposits of gypsum, many of which appear to replace the calcite wall rock. The caverns are thought to have formed where hydrogen sulphide and carbon dioxide, either as gases or in solution, moved from the oil and gas bearing Bell Canyon Formation of the Delaware Basin where they are confined by the halite beds of the Castille Formation, into the Capitan Reef (Fig. 1.17A) (Hill, 1987, 1990). Here mixing of oxygenated groundwaters from the back reef facies occurred generating sulphuric acid, which reacted with the calcite producing gypsum as a replacement product.

\[
2H^+ + SO_4^{2-} + CaCO_3 + H_2O = CaSO_4.2H_2O + CO_2
\]

Where flow was sufficient, the gypsum was leached giving cavernous porosity. Native sulphur and endellite (a kaolin mineral formed by transformation of montmorillonite under acid conditions) also characterise these processes. The hydrogen sulphide driven processes are summarised in Fig. 1.17B, which also demonstrates features of the characteristic porosity distribution. Feeder fractures guide water to a mixing zone between deep basin and oxidising waters, where cavernous porosity is best developed. The upward limit of this zone is the water table, but continued development above this level may occur by gas exhalation. The resulting caverns have a spongework pattern, unlike those formed by other processes. Details of this system are to be found in Davis (1980), Hill (1987, 1990) and Duchene & McLean (1989).

Summary

1. Deep basin waters expelled at the outflow of regional groundwater systems may contain high concentrations of hydrogen sulphide which forms sulphuric acid on mixture with oxidising groundwaters.
Fig. 1.16  A) Plan of Wind Cave showing joint controlled maze form. B) Cross-section through the southeastern flank of the Black Hills, South Dakota, showing the location of Wind Cave (modified from Bakalowicz et al., 1987).
Fig. 1.17  A) Karst development in the Guadalupe Mountains of New Mexico and West Texas. Hydrogen sulphide gas mixes with oxygenated groundwaters, the former derived from the adjacent oil field. Based on Hills, 1987, 1990). B) Cave passage dissolution by hydrogen sulphide in an oxidising mixing zone based on Carlsbad Cavern.
2. Porosity developed by this process shows strong water table control, with very large cavities, replacement gypsum and endellite clays.

3. Carbon dioxide and hydrogen sulphide may also be expelled as gases and dissolve in vadose or phreatic groundwaters to drive dissolution of carbonate.

1.1.6 Mineral Stability

Process The thermodynamic stability of high magnesium calcites is less than that of aragonite, which is some 1.5 times more soluble than low magnesium calcite. The enhanced solubility is dependent on the mole proportion of MgCO₃, which may vary between less than 5 to 30% but is typically 14%, and the ratio of magnesium to calcium in solution. Thus for 12 mole% MgCO₃ the relative solubility increases from 3.4 for a Mg:Ca ratio of 5, to 5.0 relative to calcite for a Mg:Ca ratio of 0.2, and is as high as 15.0 for this ratio and 18 mole% MgCO₃. Walter (1985) suggests that these estimates of the solubility of high magnesium calcite are too high, and that up to 12 mole% MgCO₃ the solubility is similar to that of aragonite.

For modern carbonate porewaters, Morse et al. (1985) suggested that the high reaction rate of high magnesium calcite controlled solution equilibrium, despite the relatively small proportion of this mineral in the sediment. Other workers such as Plummer et al. (1976) suggest that equilibrium with respect to aragonite is obtained. In either case, the precipitated solutions will be supersaturated with respect to calcite, which may be precipitated as a stable phase, lowering ion activities in solution and permitting further dissolution of the less stable minerals. Budd (1988) demonstrated for two islands in the Schooner Cays oolitic tidal bar belt on the Bahama Banks, that nine times more carbonate was dissolved by this conversion process than by carbon dioxide, and ten times more than by inorganic mixing. The overall calcium carbonate conversion efficiency is high (about 90%) for the freshwater lens and upper mixing zone, with some vertical transfer from the former to the latter. With increased recharge and flushing the proportion discharged from the system may however increase.

Effects Land et al. (1967) recognised a general five stage sequence of freshwater diagenesis for the Pleistocene aeolianites of Bermuda (Fig. 1.18). Initially point contact cementation occurs, followed by elimination first of high magnesium calcite, then aragonite, both of which are converted to low magnesium calcite. Harrison (1975) demonstrated that in Barbados this sequence involved the occlusion of primary porosity by calcite cements, and the development of mouldic and vuggy secondary porosity, which is commonly spatially heterogeneous. There is often no overall change in porosity, although solutional lowering of the surface may generate some autochthonous cement. There is however, a tendency to higher permeabilities on stabilisation because of the larger pore size. Stabilisation does not therefore alter the porosity distribution, which is characteristically uniform through the freshwater zone.

Controls Apparent rates of stabilisation are controlled by the relative proportion of unstable and stable carbonates present in the initial sediment. The rates reported for aragonite ooid shoals in the Bahamas are thus an order of magnitude higher than those reported for mixed mineralogy aeolianites from Bermuda (27 - 55 compared with 0.32 cm³ aragonite/m²/yr, Halley & Harris, 1979). Walter (1985, 1986) has shown that dissolution rates are also dependent on surface roughness when solutions are undersaturated with respect to aragonite and calcite, but that this dependence is reduced and mineralogy is dominant as argonite saturation is approached. The degree of leaching, which is dependent on the effective precipitation is also a major factor. Harrison (1975) reports that high magnesium calcite (but not aragonite) is still present in sediments 300ka old on the arid south coast of Barbados, while it is lost on the more humid west coast by 83ka, and aragonite by 200ka. Shedding of precipitation by a caliche crust developed during arid climates may reduce
Fig. 1.18  Sequential changes in the stabilisation of a marine carbonate sand containing aragonite and high magnesian calcite (text). I - unconsolidated sediment, II - primary skeletal grains with point contact cement, III - low Mg calcite and aragonitic grains only, IV aragonite depleted, V - stabilised limestone, low Mg calcite only.
recharge, as is the case in the Yucatan, where high magnesium calcite is still present in Pleistocene aeolianites 25 to 30ka old (Ward, 1978). In comparison, stabilisation of Holocene deposits which lack the caliche would be expected within 20 ka. Wright (1988) reports an almost identical case from the Carboniferous Limestone of South Wales.

It is important to recognise that similar incongruent dissolution/precipitation reactions may occur during later dissolution of carbonate aquifers. The most common situation is for dissolution of dolomite when the solution is saturated with respect to calcite (Wigley, 1973).

Summary

1. High magnesium calcite and aragonite are less stable than calcite.

2. Simultaneous dissolution of these minerals and precipitation of calcite may occur with high efficiency.

3. Stabilisation does not result in a change in overall porosity or its distribution, but gives an increase in permeability because of the vuggy and mouldic secondary porosity developed.

4. Stabilisation is most rapid in fine grained sediments with large surface area under a humid climate where recharge may occur unhindered.

5. At high degrees of undersaturation, grain size and surface roughness may be more important than mineralogy in controlling rates of dissolution.

1.2 HYDROLOGY OF CARBONATE TERRAINS

The circulation of water within carbonates both supplies the chemical agents driving dissolution, and removes the products of this dissolution. The hydrology of carbonate terrains is therefore a major control on both the style and distribution of dissolutional porosity. In the following account, carbonate terrains are separated into continental and oceanic domains. In the oceanic domain the density of both groundwaters and adjacent seawaters (dependent on salinity and temperature) controls the circulation and distribution of groundwaters. The transition zone between marine saline waters and freshwater is often relatively sharp. In the continental domain, elevation head controlled by uplift is most important in driving circulation, and salinity gradients within aquifers are much less steep.

1.2.1 The Continental Domain

Circulation in sediments results initially from pressure heads developed by compaction of sediments during burial. Many authors have suggested the resulting flows are important in developing secondary porosity and transporting hydrocarbons. However, numerical modelling has demonstrated that the groundwater velocities are low in such flows (0.2 cm/a), because the volume of water involved is limited to that incorporated in the pores (Bethke, 1985; Fig. 1.19). Volumes and velocities increase linearly with sedimentation rates. Flow initially is upward through the overlying sediment, but with continued subsidence, there is a tendency for lateral flow towards the basin margin, as a result of the anisotropic permeability of the layered strata.

Orogenic movements eventually cause deformation and uplift of the basin sediments, and fluid flow is then driven by the hydraulic gradient created by the slope of the water table, which is a subdued replica of the topography. Water moves from areas of high to low elevations under the influence of gravity. Unlike the compaction driven circulation which is transient, this groundwater circulation may be considered a steady state flow and is controlled by the volume of water entering the aquifer, the effective recharge.

Groundwater flow systems may be considered in terms of local, intermediate or regional scale systems with horizontal dimensions from one to hundreds of kilometres (Fig. 1.20). In areas of low relief, only regional systems develop, but where there is pronounced local
Fig. 1.19 Compaction driven flow within a subsiding basin (vertical exaggeration times 70). Fluid velocities (arrows) are shown relative to the subsiding medium (modified from Bethke, 1985).
Fig. 1.20 Conceptual model of a gravity driven fluid flow in sedimentary basins showing local, intermediate and regional flow systems. A is the recharge zone and B is the regional discharge zone.

Fig. 1.21 Effects of topography on scale of flow system. A regional flow system develops under low relief (A) but where there is more prominent relief (B), only local systems develop (modified from Freeze & Witherspoon, 1967).
topography, local systems dominate (Fig. 1.21). Many studies of karst hydrology concentrate only on the local scale, where circulation is relatively shallow, but regional scale systems may extend to considerable depth. Thus secondary porosity development is not necessarily limited to shallow depth in palaeokarst terrains, as is commonly assumed.

1.2.1a Regional Systems As shown in Fig. 1.21, in the recharge area hydraulic potential is downward, and water may move across formation boundaries and into lower units. This movement is accentuated if the underlying unit has a high permeability (Freeze & Witherspoon, 1967), for instance a fractured or karstified limestone overlain by shales. As a result of the high hydraulic gradient developed across the overlying low permeability shales, inflow to the lower unit is incremenental downgradient, giving increased groundwater flow velocities in the downstream direction. This effect is most pronounced in large groundwater flow systems. The hydrology and hydrochemistry of these outflows is sufficiently different from that of the recharge area and local to intermediate systems, that we consider them to comprise a distinct continental hydrochemical karst facies.

1.2.1b Local and Intermediate Systems The hydrology and karst style of local and intermediate scale groundwater systems is strongly controlled by the geological structure, disposition of carbonate and non-carbonate rocks and the relief (Smart & Hobbs, 1987). Together these control the distribution and nature of the recharge to the aquifer and the geometry of the internal flow paths.

Recharge may occur directly from the exposed limestone surface (authigenic recharge) or from adjacent or overlying non-carbonate rocks (allogenic recharge).

Allogenic recharge is often undersaturated with respect to calcite because it has not previously been in contact with carbonates, and it therefore has a high potential for dissolution. The type of porosity which develops is however controlled by the degree of concentration of the inflow. Where recharge is dispersed and via leakage from an overlying bed, many joints will be equally enlarged giving a maze pattern (Fig. 1.22B), and a relatively homogeneous fissure porosity. In the case of concentrated recharge, either via stream sinks at the border of the limestone area, or vadose shafts along the margins of the overlying impermeable beds, a highly heterogeneous system develops with major conduits forming from the points of recharge.

Authigenic recharge generally generates a relatively homogeneous distribution of porosity, particularly in porous or densely jointed limestones. However, in massive limestones, concentration occurs in the shallow vadose zone - the subcutaneous zone (Smart & Friederich, 1987; section 1.1.1), and a relatively heterogeneous cavernous porosity is again developed (Fig. 1.22A). Recharge may also strongly control the spatial distribution of secondary porosity. Thus in the Cretaceous Chalk of Southwest England, enhanced circulation has created zones of high transmissivity underlying the dry valley network from which recharge occurred during colder climates in the Quaternary.

Discharge from the carbonate unit is controlled by the interaction between relief and geological structure. Where relief is high, an extensive unsaturated (vadose) zone will develop in which water movement is vertical, and discharge will tend to occur from the lowest point on the carbonate outcrop. This forms the base-level for the aquifer; below this elevation the limestones are perennially saturated, but above there is a zone of dynamic storage in which water levels fluctuate on a periodic or seasonal basis in response to variations in recharge. This zone is often one of enhanced dissolutional porosity as demonstrated for the Chalk by Foster (1974).

Where regional base-level lies below the base of the limestones, freeflow or perched conditions develop, and there is little or no saturated storage. The resulting concentration of dissolution activity at the contact gives rise to laterally extensive cavern systems such
Fig. 1.22  A) Porosity distribution in a maturely karstified aquifer with a well developed subcutaneous zone and conduit flow controlled by the level of the regional water table (modified from Paloc, 1977). B) Section through a hypothetical karst aquifer showing the aquifer type and main components of recharge, storage and flow.
as Sarawak Chamber (700 by 400m), the largest chamber in the world. Such large voids are frequently unstable, and zones of brecciation will result from their collapse. Similar zones of brecciation are frequently reported in palaeokarst literature, and have high porosity (Kerans, 1988), although their formation is often problematic. Perching may also occur in impure limestones containing shales and sandstones.

Circulation of Groundwater below regional base-level may occur in vertically extensive carbonate aquifers, either in the perenially saturated unconfined zone or in the confined zone where overlying impermeable rocks restrict upward discharge of water (Fig. 1.22B). Circulation in confined aquifers is often restricted away from outcrop, where flow is part of the intermediate or regional scale systems, and isolation from both vadose inflows and gas exchange has important geochemical effects.

Depth of circulation in the unconfined zone is often controlled by geological structure in massive well-cemented carbonates, as secondary porosity is developed preferentially on laterally continuous bedding planes compared to discontinuous joints (Ford, 1988). Thus in areas of low regional dip, circulation below base-level may be limited to tens of metres as in Northwest Yorkshire (Fig. 1.23). In steeply dipping limestones, water is led down-dip to depth, and must follow joints to gain stratigraphic height, and discharge at the limestone margin. This is the pattern of cave development on the Mendips (Fig. 1.23). However, as shown by Ford and Ewers (1978), with increasing frequency of fractures, the depth at which caverns develop is reduced, and eventually approximates the position of the water table (Fig. 1.24).

Storage. It is important to remember that although cavernous porosity may be responsible for the majority of circulation in karstified aquifers, caverns contain only a small proportion of the total water in storage. Much of the storage is in fractures and fissures with a diffuse flow regime in the limestone blocks between conduits (Fig. 1.22). Interactions between the conduit and diffuse flow components are complex, but water may move from the conduits to diffuse storage at times of recharge (for instance transporting dissolved organic matter into the fissures for later oxidation, section 1.1.2), with subsequent reversal at low flow. This pumping action is important in developing the diffuse flow porosity, giving a dual porosity system, which may create significant problems for oil recovery.

As described in section 1.1.1, dissolution rates in carbonate areas are controlled dominantly by effective precipitation or runoff (precipitation minus actual evapotranspiration). The permeability of cap rock beds overlying the limestone is also important, where permeabilities are low, shedding of some precipitation may occur. The development of thick clay residual deposits may similarly restrict recharge, as may laterally continuous caliches developed in the soil and as a crust at the bed rock surface. In recently exposed carbonates with intergranular flow, caliche horizons cause lateral concentration of recharge, developing vadose shafts even though intergranular porosity is high, and giving a heterogeneous porosity distribution. The restricted recharge below such caliches also explains the anomalous persistence of high magnesium calcite and aragonite in Pleistocene aeolianites in the Yucatan (Ward, 1978).

Summary

1. Steady state flows developed on uplift as a result of elevation differences are much more important than expulsion of waters during compaction for development of dissolutional porosity.

2. Regional groundwater systems dominate in areas of low relief, focusing large volumes of water towards regional discharge zones, and giving circulation at depth.

3. Recharge type is significant in controlling the heterogeneity of dissolutional porosity generated in carbonate aquifers.

4. The elevation of the regional base-level with respect to the base and top of
Fig. 1.23 Effect of geological structure on the depth of karstification in the saturated zone. Note that the depth of the unsaturated zone (controlled by relief) is also significant for the total porosity distribution.
the carbonate units is critical in controlling the distribution of cavernous porosity.

5. Geological structure and the density of fracture porosity is important in controlling the depth to which cavernous porosity is developed.

6. Many carbonate aquifers will develop dual porosity systems with cavernous, fracture, fissure or intergranular flow.

7. Recharge may be restricted by caprocks or surficial materials such as caliche crusts overlying the limestones.

1.2.2 Oceanic Domain

Two types of oceanic karst can be recognised where carbonate aquifers are in continuity with the sea:

1. Margins of continental carbonate terrains where sea water intrudes from the adjacent oceans. These frequently comprise older rocks, characterised by well developed fracture and karstic porosity, as documented in Greece (Stringfield & LeGrand, 1969), and have a significant component of allogenic recharge (Back et al., 1984).

2. Isolated carbonate build-ups in the form of platforms or atolls completely underlain by saline water. These are often composed of relatively young carbonates with high intergranular porosity, although following cementation cavernous porosity may occur. Emphasis is here placed on the hydrological regime of the isolated build-ups.

Density is the major control on the distribution and movement of groundwaters in oceanic karst systems. The density contrasts between fresh recharge water and seawater results in the establishment of three distinct hydrological zones: the freshwater lens, which floats on the underlying saline water and is separated from it by a transition or mixing zone of variable intermediate salinity.

1.2.2a The Freshwater Lens In oceanic islands the less dense freshwater forms a lens shaped body, while at continental margins a freshwater wedge is developed. The depth of the freshwater body is controlled primarily by hydrostatic equilibrium between the fresh and saline water, as described by the Ghyben-Herzberg approximation. This defines the depth to the base of the lens below sealevel ($Z_s$) as a function of the elevation of the water table above sealevel ($Z_w$), assuming the fluids are immiscible and exist under steady state conditions in a homogeneous unconfined aquifer.

$$Z_s = \frac{\rho_f}{\rho_s - \rho_f} Z_w$$

Thus for a freshwater lens ($\rho_f = 1.000$ kg/L) overlying groundwater of seawater density ($\rho_s = 1.025$ kg/L), the depth of the lens is about 40 times the freshwater head (Fig. 1.25). However, in most real situations the Ghyben-Herzberg relationship provides only an approximation as a result of the miscible behaviour of fresh and saline groundwaters (violating the assumption of hydrostatic equilibrium) and the active circulation of the groundwaters (generating non-steady state conditions). The size and shape of the freshwater lens is dependent upon the balance between recharge, which is climatically controlled, and discharge, which depends on the bedrock permeability and aquifer dimensions.

The lens is recharged via percolation of rainfall through the unsaturated zone, although, where the unsaturated zone is thin, evapotranspiration from the water table will reduce the effective recharge. The thickness of the lens will thus reflect spatial variations in the effective recharge, and in extreme cases the lens may be segmented, as occurs with inland ponds and marshes on San Salvador Island, Bahamas (Davis & Johnson, 1989).

The width of the island provides an effective limit to the size of the lens developed, with an estimated minimum width of 400m necessary to support a freshwater lens. Budd and Vacher (1991) estimate that the lens thickness in
Fig. 1.24 Effect of primary opening frequency on the style and depth of cave development. Fracture frequency low (1) to high (4). (Modified from Ford and Ewers, 1978).

Fig. 1.25 Hypothetical cross-section showing the transition zone (marked as the zone of dispersion) and flow patterns in a homogeneous coastal aquifer wedge (modified from Cooper, 1959).
modern carbonate islands ranges from 0.2% of island width, with 1% being the most representative value. Because of the difficulty of estimating precisely the elevation of the water table, especially for the palaeo-islands, island width provides a useful surrogate measure for predicting lens thickness.

Variations in hydraulic conductivity, both horizontally and vertically, provide an important control on the shape of the freshwater lens. In Pleistocene aeolianites of Bermuda, Vacher (1978) found that a thicker lens was developed in less well karstified younger deposits, resulting in highly asymmetric lens distribution. Similarly in the Bahamas, more karstified, higher permeability limestones occurring at depth, effectively truncate the base of the freshwater lens (Cant & Weech, 1986). Under confined conditions the freshwater lens may extend a considerable distance from the shoreline as observed in the coastal wedge of northern Florida (Johnson, 1983), although such cases are rare for oceanic islands.

Maximum lens development will occur under conditions of high rainfall and low evapotranspiration on a large, circular island with high relief, low permeability and dominated by intergranular flow.

1.2.2b The Transition or Mixing Zone The interface between fresh and saline groundwaters is rarely a sharp boundary between two immiscible fluids, but forms a zone of intermediate salinity termed the transition or mixing zone. Under static conditions the freshwater lens may extend a considerable distance from the shoreline as observed in the coastal wedge of northern Florida (Johnson, 1983), although such cases are rare for oceanic islands.

Maximum lens development will occur under conditions of high rainfall and low evapotranspiration on a large, circular island with high relief, low permeability and dominated by intergranular flow.

Deep Saline Groundwater System Recent studies suggest that saline waters may play an important role in carbonate diagenesis (e.g. Saller, 1984; Land, 1985), generating karstic porosity by dissolution and the potential for fracture porosity by cementation and dolomitisation.

Three forces are capable of driving regional scale flow in the subsurface of carbonate build-ups, pressure, density and temperature.

Pressure - Elevation Head A difference in water surface elevation across an island or bank can drive flow in the saline system (Fig. 1.27). This may be developed on a semi-diurnal basis in response to variations in the rates of tidal ebb and flow between the restricted bank or lagoon environment and the open ocean (Fig. 1.28A). These short term tidally driven flows are important in supratidal environments (Fig. 1.28B) and may be
Fig. 1.26 Exponential reduction in the transition zone thickness with distance from the coastal (•) or creek (○) margin. Group A includes sites less than 500m from tidal creeks, while group B sites occur in zones of very high permeability associated with relic tidal creeks (Whitaker, 1991).
Fig. 1.27 Schematic diagrams illustrating the processes that drive circulation of saline water in carbonate platforms. No stipple = freshwater; dashed lines = transition zone; light stipple = lower density saline water; heavy stipple = higher density saline water. From Whitaker & Smart, 1990.
responsible for cementing atoll and platform margins (Aissoui et al., 1986). Longer term head differences due to wind and storm controlled variation in sea surface elevation on leeward and windward shores are also observed. On Davis Reef, Australia, even in the absence of major barriers and enclosed lagoons the reef systems generate head differences which are wind and tidally induced (Oberdorfer & Buddmeier, 1985). Although regional scale circulation systems have not been generally recognised, recent work on North Andros, Bahamas, indicates an easterly flow of cold seawater at depth beneath the Great Bahama Bank, probably related to head differences generated by the Florida Current (Whitaker & Smart, 1990).

Density - Reflux Ocean water concentrated by evaporation on the shallow banks moves downwards, displacing less dense fluids, until it reaches a depth where its density is equivalent to that of the adjacent ocean, when lateral flow occurs (Fig. 1.27). This reflux circulation system was originally proposed for very restricted environments such as the hypersaline lakes of Bonaire (Lucia, 1968). However, Simms (1984) demonstrated that reflux can occur where waters have only slightly elevated salinity (minimum enrichment of 2 to 5 %o), such as occurs over large areas of the Great Bahama Bank today (Queen, 1978). These elevated salinity waters can be traced beneath North Andros Island, and also discharge from the platform margins at depth (Whitaker & Smart, 1990).

Extensive areas with shallow seas (determined by the position of sea level), and high evaporation rates are necessary for reflux to occur on a large scale. Reflux plumes are essentially cylindrical, although if long-lived, large stratiform bodies form as the plumes change position.

Density - Buoyant Circulation Saline waters flow into the interior of the island in response to buoyant circulation in the transition zone (Fig. 1.27 and see section 1.2.2a). This flow system is characterised by the discharge of large volumes of brackish water at the coast, as reported off the coast of southeast Florida by Kohout (1960).

While there are many studies of the transition zone as a geochemical environment for dissolution, and as a pump for circulating mixed waters, the role of buoyant circulation in driving saline flow is rarely considered. In oceanic islands buoyant circulation is only effective on a large scale during periods of low sea level since subaerial exposure is required to develop an extensive and thick freshwater lens. However, it is likely to be very well developed at continental margins due to increased hydrological activity in the transition zone, and in ramps buoyant circulation is the dominant mechanism driving saline groundwater circulation. This mechanism has been implicated in the formation of large volumes of dolomite beneath the Little Bahama Bank by normal seawater (Vahrenkamp, 1988). However, the torroidal shape of the zone of influence of buoyant circulation is incompatible with the wedge or lensoidal dolomite bodies.

Temperature - Convective Circulation This process may be initiated by a horizontal density gradient between cold ocean water surrounding the carbonate build-up, and groundwaters within which are warmed by the geothermal heat flux. In Fig. 1.27, cold dense water is drawn into the platform at great depth, where it is heated, and rises to discharge at submarine springs along the platform margin. Convective flow is enhanced by the presence of a thermal anomaly, such as the volcanic pedestal which drives circulation beneath Niue Atoll in the Pacific (Aaharon et al., 1987).

Kohout first recognised thermal convection in the Florida Peninsula (Fig. 1.29), and suggested a relation to dolomitisation at depth (Kohout, et al., 1977). Convective flow cells can be identified from the temperature distribution, with negative geothermal gradients persisting to considerable depth near the coast, and from warm, geochemically evolved springs on the platform margin (Fanning et al., 1981). The retrograde solubility of carbonates (see section 1.1.4) creates the potential for dissolution by water heated in the
Fig. 1.28 A) Head differential resulting from the frictional retardation of semi-diurnal tides on the shallow carbonate banks (modified from Matthews, 1974). B) Active pumping in supratidal carbonates at Sugarloaf Key, Florida (modified from Carbello et al., 1987).
Fig. 1.29 Kohout’s (1960) model of thermal circulation in the artesian aquifer of southern Florida (Miami region).
centre of the build-up, and reprecipitated at the margins, where cooling occurs.

**Implications**

From the above it can be seen that there are opportunities for significant karstic porosity formation in carbonate build-ups related to even minor sea-level changes. Such porosity formation, linked with the range of possible flow mechanisms within the systems, can create both a 'plumbing system' and a flux for subsequent diagenesis. The importance of these processes in the diagenesis of ancient buildups is graphically illustrated by the recent studies of the late Triassic - early Jurassic platforms of Austria (Mazzullo et al., 1990). In these platform limestones, zones up to 180m thick show extensive limestone dissolution and early cementation by marine cements. Low stand phases produced both mixing and meteoric dissolution, followed by extensive cementation during highstands, from marine or marine-derived fluids circulating at depths of up to 180m.

**Summary**

1. The oceanic domain comprises the marine margins of continental carbonate terrains, and isolated carbonate build-ups.

2. A hydrostatic equilibrium between fresh and underlying saline (dense) water approximated by the Ghyben-Hersberg relationship, controls the thickness of the freshwater phreatic zone.

3. The volume of the freshwater phreatic zone is controlled by effective recharge, aquifer size and aquifer permeability.

4. The thickness of the transition zone between fresh and saline water is controlled by seasonality of recharge, permeability, and distance from the coast or discharging creeks.

5. Large volumes of saline water may move beneath the freshwater lens, but their geochemical effects are poorly understood. Buoyant circulation is not thought to be a major driving process.

6. Elevation heads developed in response to tides, winds, storms and regional ocean currents may drive saline water circulation across platforms.

7. Density reflux as a result of evaporative concentration on the banks may give cylindrical circulation plumes within platforms.

8. Convective circulation may result from the temperature contrast between cold ocean water and platform groundwaters, giving inflow of marine water at depth.

### 1.3 POROSITY IN CARBONATES AND THE ROLE OF DISSOLUTION

#### 1.3.1 Recent Carbonates

For most recent carbonate sediments, porosity and permeability are inversely related (Fig. 1.30A), and are controlled by the proportion of the sub-62μm grains. Stabilisation of the unstable phases high magnesium calcite and aragonite to low magnesium calcite (section 1.1.6) causes little change in overall porosity, but the preferential development of cements within the smaller pores, and more importantly the development of larger fabric selective mouldic and vuggy porosity gives an increase in permeability, because flow is proportional to the third or fourth power of the pore diameter. Pittman (1974) has for instance shown this effect in Pleistocene corals from Barbados (Fig. 1.30B). The porosity developed on exposure of young carbonates is thus typically spatially variable on both large and small scales, with larger isolated voids (vugs and moulds) separated by finer pores, and zones of more intense solution surrounded by less altered rock as in Fig. 1.31 (Harrison, 1975).

With time, there may be additional development of karstic porosity, but overall cementation and porosity occlusion is dominant giving a considerable range of porosity values (Fig. 1.32). This high degree of heterogeneity is characteristic of karstified limestones. Further diagenetic changes and compaction occur with
Fig. 1.30  A) Relationship between porosity and permeability for Holocene carbonate sediments (modified from Enos & Sawatsky, 1981).
B) Relationship between porosity and permeability for stabilised (calcitic) and nonstabilised (aragonitic) corals (modified from Pittman, 1974).
Fig. 1.31 Effect of burial on the type of dissolitional porosity development in carbonates. $O =$ porosity, $K =$ bulk permeability, $H =$ heterogeneity, $C =$ connectivity.

Fig. 1.32 Porosity versus time plot for three oolitic sands. The Joulter's Cay oolite has been exposed for $10^3$ years whereas the Miami Oolite is Pleistocene in age. The upper curve represents the development of karstic porosity while the lower one represents porosity loss by cementation. Based on Halley & Evans, 1983; Evans & Ginsburg, 1987.
Fig. 1.33 Relationship between porosity and depth for limestones and dolomites in South Florida (see text).

Fig. 1.34 Secondary and primary porosity, pore-size and permeability of carbonate rocks. Contour values of $K$, the theoretical permeability, are based upon the assumption that the rock behaves as a bundle of straight, parallel capillary tubes. Values shown underlined indicate the general range of permeability found in the lithology concerned. Double underlining indicate total porosity, single underlining indicates primary porosity (modified from Smith et al., 1976).
Fig. 1.35  A) Effect of void integration in controlling permeability in fissured limestones (modified from Sandlein & Palmquist, 1977).  
B) Relationship between specific yield (porosity drained under gravity) and permeability for boreholes in the Carboniferous Limestone of the Mendip Hills (modified from Hobbs & Smart, 1988).
burial, and in the Florida area amount to a halving of porosity every 1,740 m of burial, a pattern independent of age (Schmoker and Halley, 1982; Fig. 1.33). Note that for dolomites, porosity is less than for limestones at shallow depth, suggesting that replacement destroys porosity (or that finer limestones are preferentially dolomitised). Below 1,700 m this pattern is reversed, with dolomites being more resistant to burial. This is clearly of considerable significance when considering reservoir potential.

1.3.2 Effects of Re-exposure After Burial

Many limestones have systems of joints and bedding planes, developed as a result of a reduction in lithostatic pressure on re-exposure after burial. Where there has been substantial occlusion of pre-burial porosity, these mechanical, secondary openings are the major type of porosity present (Fig. 1.31). The majority of caves are formed along bedding planes, which are more laterally continuous than joints, which often die out laterally and vertically. Very few caves are known to develop directly from intergranular porosity, although those associated with transition zones may be an exception. Thus, in comparison with intergranular porosity, secondary mechanically derived porosity has a major influence on the circulation of fluids, and the development of dissolutional porosity because of the large effective pore-diameter and lateral continuity. The permeability increase is also more significant than that in porosity (Fig. 1.34), because of the diameter/flow power law. Porosity and permeability are generally positively correlated (Fig. 1.34B), a reversal of the trend for carbonate sediments described above.

Fracture development is not however limited to ancient deeply buried carbonates. In many carbonate platforms, networks of fractures may develop parallel to the margins, either by differential compaction (Playford, 1984), or as a result of unconfined lithostatic pressure along the platform margin (Smart et al., 1988b). Thus Dougherty et al. (1986) have mapped neotectonic joints in carbonates less than 150 ka in age on North Andros, Bahamas. On South Andros, dissolutional development along these bank marginal fracture systems has produced substantial cavernous porosity, which has a significant effect on the modern groundwater hydrology.

1.3.3 Nature of Dissolutional Porosity

Occlusion of pre-burial porosity in carbonates is often incomplete. Thus many modern karstified limestones show a mixture of recent dissolutional porosity controlled by mechanical openings, and pre-burial 'intergranular' porosity (Table 1.2). For the Carboniferous Limestone, primary porosity is very low and total effective porosity in the aquifer is dominated by fissures (solutionally enlarged fractures). In contrast, while the specific yield of the Chalk is dominated by fracture porosity, and is similar to that of the Carboniferous Limestone, the total porosity of 30 to 40% is dominated by very fine intergranular and intra-particle porosity. The Jurassic Great Oolite occupies an intermediate position with a high fracture and intergranular porosity. Thus many carbonate aquifers with secondary dissolutional porosity developed either pre- or post- burial, exhibit a dual porosity behaviour, with fracture controlled or vuggy porosity, giving high transmission rates, but substantial storage in small diameter inter- and intra-particle porosity.

Two additional characteristics of dissolutional porosity are of concern to carbonate reservoir engineers: heterogeneity and connectivity. As discussed above, fracture systems give the potential for high void integration (connectivity), but this must be developed by dissolution. Thus in immature karst aquifers, void integration may be low (Fig. 1.35), and hydrocarbons enclosed in the pores are not readily extracted. As dissolution proceeds, void integration increases, thus the maturity of palaeokarst is of considerable importance (see section 1.3.4). High void integration may also result from a high density of
Fig. 1.36  
A) Hydraulic conductivity data (A) for the Carboniferous Limestone, Mendip Hills (see Hobbs & Smart, 1988). For comparison, specific yield (B) for a fractured dolomite aquifer from Illinois is added (see Csallany & Walton, 1963).

B) Specific capacity (yield per unit drawdown per unit aquifer permeability) for boreholes in valleys and uplands in a dolomite aquifer (modified from Csallany & Walton, 1963).
initiating fractures, for instance in thin bedded peritidal limestones.

Karst aquifers are notoriously heterogeneous, at both the void and regional scales. Intergranular permeability is generally highly uniform, but, with increasing dissolution, mouldic and vuggy porosity give an increase in small scale heterogeneity. However, where post-burial karstification occurs, very high heterogeneity may result. Fractures (width mm) are solutionally modified to give nets of fissures (width cm), both being homogeneous on a scale of tens of metres. Where concentrated recharge occurs, and in very massively bedded carbonates, cavernous porosity develops. Large flows are accommodated in single high capacity voids of limited size, and the probability of intersection during drilling is low. For the cavernous Carboniferous Limestone, hydraulic conductivity measured by borehole tests varies over 5 orders of magnitude, compared to only 2 orders of magnitude for a fracture flow aquifer. Thus cavernous porosity is intrinsically very heterogeneous and represents a major problem for development.

On a regional scale, even for a relatively homogeneous fracture flow aquifer such as the Great Oolite of the Cotswolds, zones of preferential dissolutional development may occur. Thus valleys frequently have greater recharge (and hence dissolution) than uplands, and higher well yields are observed (Fig. 1.35). An understanding, both of the hydrology and dissolutional processes is necessary to predict development of porosity in karstified reservoirs.

The major controls on the development of dissolutional porosity in carbonates are therefore:- whether karstification was pre- or post-burial; the nature of the fracture system, and the hydrological and geochemical processes operative.

Summary

1. Dissolutional porosity developed in carbonate rocks prior to burial is often vuggy and mouldic in character, and is spatially variable.

2. Burial reduces porosity substantially, but gives the potential for fracture development on subsequent exposure.

3. Fracture based porosity gives the potential for high permeability in post-burial carbonate rocks, the actual magnitude of which is controlled largely by the extent of subsequent dissolution.

4. Many carbonate rocks have dual porosity controlled by post-burial fractures and pre-burial intergranular porosity.

5. Void integration is low initially but increases with the extent of bedrock dissolution.

6. Karstified aquifers exhibit substantial heterogeneity in permeability at both void scale (intergranular < mouldic and vuggy < fracture < fissure << cavernous), and at the regional scale.

7. Understanding the hydrological and dissolutional processes responsible for porosity generation is essential for effective prediction and development of reservoirs in carbonate rocks.

1.3.4 Evolution of Porosity Through Time

A key concept in understanding karst/palaeokarst porosity is that dissolutional porosity develops progressively with time. For example, an ancient limestone uplifted to the surface will have fractures less than 1mm wide (Fig. 1.37). Circulation is initially limited and dissolution slow, but at a diameter of 5 to 10mm three critical thresholds are exceeded:

1. flow changes from laminar to turbulent.

2. penetration of significantly undersaturated waters significantly increases solution rates.

3. removal of sediment by the turbulent flow prevents blockage and the build-up of protective surface residues.
Fig. 1.37 Schematic rate curve for the evolutionary history of a single conduit. The time scales are drawn from a combination of field evidence and geochemical calculation (modified from White, 1988).

Fig. 1.38 Change of mean elevation with time for karst terrains with different rates of continuous uplift and denudation, demonstrating the effects of elevation and uplift on time available for karstification. 1. Large initial uplift, 2 small initial uplift. Total relief (mean elevation - fluvial (incision) is shaded.
Thereafter enlargement of the initial opening proceeds rapidly to give a cave passage, although where recharge is uniform or discharge is limited by competing flow routes, the onset of this stage appears to be deferred indefinitely to give a fissure flow aquifer. Ewers has demonstrated, using salt block models in the laboratory, that the development of cavernous porosity has a major effect on flows within the aquifer. Inputs are redirected to the pioneer route, and an integrated network develops.

However Fig. 1.37 also demonstrates that development cannot be maintained indefinitely, and that destruction of the cavernous porosity may also occur. Three destructive agents are important:

1. The mechanical competence of the host rock may be exceeded and collapse can occur.

2. The passage may be captured to a new lower route, due to the change in base level or development of a hydraulically more efficient route through the limestone along with increasingly frequent dissolusional openings.

3. The conduit may be eliminated when surface lowering reaches it.

These retrograde processes, together with sedimentation causing blocking, abandonment and burial, are termed dekarstification, and may be of considerable importance in understanding the nature of buried palaeokarst reservoirs.

1.3.4A Karstification - Porosity Generation

The total dissolutional porosity can be considered simply as a product of the subsurface dissolution rate and exposure duration. The total porosity created is, however, distributed through the effective aquifer thickness available, the distribution being controlled by the hydrological and geological processes active:

\[ + \Delta \phi = f \left( \frac{S_{\text{Sub}} - t}{d} \right) \]

where \( \Delta \phi \) = change in porosity; \( S \) = dissolution rate, \( t \) = exposure duration and \( d \) = aquifer thickness.

Exposure duration is a function of the initial relief, which may be controlled by either a fall in sea level or tectonic uplift of the area, and the relative rate of continuing uplift (or sea level fall) compared to the rate of surface lowering. Thus:

\[ t = f(t_0 + r_0 \left( U/S_{\text{Sur}} \right)) \]

where \( r_0 \) = initial relief, \( U \) = rate of continuous uplift, and \( S_{\text{Sur}} \) = rate of surface lowering.

In a basin history context for a carbonate platform, small scale changes in sea level give frequent but short intervals of exposure and karstification, as may be seen in many peritidal sequences (Fig. 1.41). Longer term Milankovich cycles give more prolonged exposure, with complete mineralogical stabilisation and significant secondary porosity development. Finally, major changes in eustatic sea level at the time scale of the Vail cycles give regional unconformities and evolution of maturely karsified terrains (see Section 2.4.3).

Kars landscapes may be mapped into a continuum, ranging from those with initial uplift followed by denudation (the classical Davisian cycle of erosion), to those where denudation and uplift are essentially balanced (the dynamic equilibrium landscape of Hack), and finally those where relief is not limited by denudation, which is comparatively slow, but by lithostatic effects (the orogenic zone) (Fig. 1.38). The bewildering variety of tower and cone karst landscapes in Southern China can only be explained using these concepts (Smart et al., 1987). However it is also important to realise that fluvial incision is generally more rapid than surface lowering, thus even under dynamic equilibrium the base level controlling the groundwater hydrology of karst areas may fall. Thus it is normal to see sequences of abandoned cave-levels in modern karst areas (Fig. 1.39), and progressively with time each volume of rock will pass through the different hydrological zones, from deep to shallow.
Fig. 1.39 Passage levels in Mammoth Cave, Kentucky. A - late Tertiary canyon passages partly filled with clastic sediments; B - early Pleistocene tubular passages; C - mid-Pleistocene tubular passages which correlate with remnants of a major terrace in nearby river valleys; D - late Pleistocene passages, now partly flooded by a Wisconsinian rise in base-level; E - minor canyons and shafts formed in the vadose zone. Sedimentation of the flooded passages by backflooding is currently occluding karstic porosity.
Table 1.2  Specific yield (gravitationally drained porosity) and total porosity for carbonate aquifers in the United Kingdom. Figures are percentages.

<table>
<thead>
<tr>
<th>Aquifer Type</th>
<th>Specific Yield</th>
<th>Total Porosity</th>
</tr>
</thead>
<tbody>
<tr>
<td>Carboniferous Limestone</td>
<td>0.5 - 1.0</td>
<td>0.5 - 1.0</td>
</tr>
<tr>
<td>Cretaceous Chalk</td>
<td>1.0 - 1.5</td>
<td>30 - 40</td>
</tr>
<tr>
<td>Jurassic Great Oolite</td>
<td>2.0 - 3.0</td>
<td>7 - 15</td>
</tr>
</tbody>
</table>

Fig. 1.40  Change in total porosity with time for uplifted carbonate terrains as a function of rates of creation and destruction. Bar shows effective porosity at time $t$. 
phreatic, to vadose and finally the subcutaneous zone.

Solution rate is controlled predominantly by the regional climate; temperature controlling reaction rate and soil PCO₂, and effective rainfall controlling soil PCO₂ and runoff. Mineralogical stability, hydrological function and domain are also important. Thus much longer will be required to develop a particular level of porosity in a semi-arid environment (recognised by the caliche facies of Esteban & Klappa, 1983), than in a humid area ('bare' karst facies), although differences between soil-covered and non soil-covered terrains would be expected.

1.3.4B Dekarstification - Porosity Destruction

Two processes of porosity reduction are important, surficial erosion and sedimentary infill. Collapse and brecciation are actually porosity creation processes even though they involve elimination of cavernous porosity, and are of considerable importance in some oilfields (Kerans, 1988). Surficial erosion becomes of increasing importance as relief is lowered and the potential energy for transport reduces. Other changes such as climate or regional topography may also shed more insoluble material onto the carbonate terrain, initially mantling it, and finally choking the cavernous porosity. This process is widely documented from palaeokarst terrains in the geological record, but has received little attention in geomorphology. Examples however, include backflooding and sedimentation on the Central Kentucky karst associated with aggradation of the river which controls base level in the karst, and increased sediment production related to soil erosion in the tobacco and maize crops grown in this area.

Porosity elimination may therefore be written:

\[-\Delta \phi = f \left( \frac{S_{\text{surf}} \cdot t}{r} + \frac{(I_s + I_c) \cdot t}{d} \right)\]

Where: \(\Delta \phi\) = change in porosity, \(r\) = relief, \(t\) = exposure time, \(I_s\) = elastic sedimentary fill rate, \(I_c\) = cementation rate and \(d\) = aquifer thickness.

The general form of porosity evolution and destruction for a carbonate terrain experiencing initial uplift only is shown in Fig. 1.40. It is clear that both young and senile karst landscapes have limited porosity. That in the latter may however be well interconnected compared to that in the former, depending on the degree of sedimentation. Between these end members, the porosity of karst terrains reaches a maximum. Furthermore, if uplift and denudation are balanced (Fig. 1.39), this porosity will persist through time.

Summary

1. Porosity in palaeokarst terrains may be generated by karstification and eliminated by dekarstification.

2. The extent of karstification is controlled by the rate of dissolution and duration of exposure. The latter is dependent on initial relief and the relative rates of denudation and continued uplift.

3. The extent of dekarstification is controlled by relief and the allogenic sedimentation rate.

4. Juvenile and senile palaeokarst landscapes have limited porosity, but mature terrains with extant relief have higher porosity.

1.4 SYNOPSIS: KARST 'FACIES'

It is clear from the discussions above that there are common associations of hydrological and geochemical processes which give rise to particular styles of karst development and porosity distribution. We term these styles 'karst facies'. Three facies are considered below, in the oceanic domain marine-mixing systems, and in the continental domain local meteoric systems and regional basin 'outflows'. Switching between domains is controlled by the relative magnitude of sealevel change (Fig. 1.41). Small amplitude Milankovich driven cycles associated with an ice-free world produce only brief emergence.
Fig. 1.41 Influence of sealevel fluctuation on the development of different karst regimes. The marine domain has waters of marine origin where little karst development takes place. The oceanic domain is characterised by diagenetically immature carbonates with unstable mineralogies, high porosities and limited fracture development. Relief on exposure will be low. Meteoric and mixing dissolution processes will be important.
recorded as explosive surfaces in peritidal sequences, and karstification is unimportant (the marine domain). Larger amplitude Milankovitch cycles cause significant exposure with karst development in the oceanic domain. Finally, major changes in eustatic sea level and tectonism at the time scale of the 'Vail' cycles give bank abandonment, the development of regional unconformities and evolution of maturely karsted terrains in the continental domain.

1.4.1 Local Meteoric Systems

Local meteoric systems comprise the most familiar karst systems, and are the predominant type represented in both modern karst areas and in the fossil record. Groundwater circulation is driven by elevation head differentials generated by uplift, and maintained by recharge over the outcrop of the exposed carbonates. Water flow is initially predominantly vertical in the vadose zone (unsaturated) above the water table, giving markedly anisotropic and heterogeneous porosity development. In areas of high topographic relief, thick vadose zones develop, and the volumetric rate of porosity development is lower than when dissolution is concentrated into a vadose zone of more limited extent.

In the saturated zone, below the water table, the porosity generated is more uniform, and despite the predominantly horizontal flow, is less isotropic. The position of the water table is controlled by base-level, the lowest outlet point from the carbonates. In karst areas where the base of the limestone is above the base-level, horizontal flow is concentrated at the underlying contact, forming major cavernous porosity. This is often accompanied by extensive collapse of the overlying beds, producing a very high net porosity.

Cavernous porosity is usually more limited in extent and its distribution highly heterogeneous. It develops preferentially where there is concentrated recharge to the limestones from impermeable rocks, either directly overlying the limestones or forming adjacent uplands. Such clastic sequences may also provide a source of sediments, which, when transported into the limestones, infill dissolutional porosity. This porosity occlusion increases with time, and is particularly associated with senile karsts of low relief.

The major geochemical process causing dissolution in local meteoric systems is soil carbon dioxide, which generates carbonic acid. The presence of a permeable, insoluble soil cover and elevated temperature, giving high biological production and organic decomposition, therefore increases dissolution rates. Dissolution rates in bare karst areas are comparatively low. Much of the dissolutional potential of surface-derived carbonic acid is dissipated at shallow depth in surface lowering and development of a high porosity zone in the upper 5m of the limestones: the subcutaneous zone. Dissolution decreases exponentially below this depth, but is maintained by limited mixing corrosion and oxidation of organic matter giving reducing conditions at depth.

The total porosity generated is dependent primarily on the effective recharge (high in a humid climate), and duration of exposure. In arid and semi-arid climates significant redeposition of dissolved carbonates as caliches and cements at shallow depth may retard porosity development at depth by reducing groundwater circulation.

Finally the structure and nature of the carbonate provides a control on the style of porosity evolution. In immature carbonates which have not suffered deep burial, intergranular porosity is open and significant circulation occurs via these routes. In re-exposed carbonates elevated by tectonic processes, secondary fractures and bedding planes, which become more frequent towards the surface, are the predominant openings guiding water movement. Where these are frequent, dissolution is uniformly distributed, producing a homogeneous, high permeability fissure karst. Where beds are massive, flow concentration occurs leading to the development of closed depressions and conduit flow. The aquifer porosity is thus cavernous and highly heterogeneous.

The best reservoirs have:
1. Prolonged karstification at major unconformity
2. Moderate relief
3. No underlying or adjacent impermeable beds
4. Humid climate
5. Well developed soil cover
6. High fracture density
7. Moderate to low intergranular porosity
8. Pure carbonates.

1.4.2 Regional Outflow Systems

In regional groundwater flow systems, meteoric water from an extensive basin area is concentrated into carbonate units which provide the major transmission route for groundwater flow. When these units intersect the surface, an extensive zone of outflow develops. Groundwater flow is dominantly upward, and independent of local relief, which is generally low. Dissolution is driven by cooling of upward moving waters which have been geothermally heated during regional circulation, and is relatively uniform with depth. Flow is concentrated into higher permeability units, but is predominantly diffuse, following all fractures and bedding planes equally. These processes generate a characteristic network maze porosity style, which has both high permeability and low heterogeneity. Trace metal mineralisation may occur as, for example, Mississippi Valley Type deposits with which these systems are often associated.

Basin outflows may also entrain carbonic acid and, more importantly, hydrogen sulphide derived from expulsion of gases from thermally driven decarboxylation of hydrocarbons. Dissolution occurs in a limited zone near the surface (where oxidation takes place), and extensive cavernous porosity may be generated. This is often associated with precipitation of carbonate and sulphate mineral phases. Regional outflow system development is dependent more on basin area, topography and structure to a lesser extent than climate, but duration of exposure is again important. A further contrast with local meteoric systems is that occlusion of porosity by sedimentation will not occur in the upward groundwater flow, although chemical cementation may have a similar but lesser effect accumulating with time.

The best reservoirs have:
1. Prolonged exposure
2. Low topography
3. Concentration of basinal groundwater flow
4. Carbonates beneath less permeable beds
5. High fracture density
6. High geothermal heat flux

1.4.3 Marine Mixing Systems

Marine mixing facies are developed at the oceanic margins of continental carbonate terrains and in isolated carbonate islands/build-ups. These environments tend to be characterised by active carbonate deposition and thus diagenetic processes operate on a substrate which may not initially be stabilised. Although stabilisation need not increase overall porosity, the reduction in isolated primary porosity is accompanied, and often exceeded, by an increase in permeability. Stabilisation rate is controlled by internal factors, including the composition and proportion of unstable mineralologies and sediment surface area; and external conditions, including effective precipitation and extent of surface sealing, for instance by caliche.

Once stabilisation is complete, dissolution processes in the vadose and freshwater phreatic zones are essentially those occurring in local meteoric systems. However, the distinguishing feature of marine mixing facies is the contrast in density between adjacent bodies of fresh and saline water. These contrasts, in addition to hydraulic head, drive groundwater circulation, while mixing between different water bodies represents an additional dissolutional drive, concentrated in particular in the transition zone.

Porosity creation is characteristically concentrated along laterally continuous horizontal or lenticular zones, with maximum development at the exposure surface, and beneath in the vadose zone and at the water table. Porosity in the phreatic zone, which is dominantly vuggy and
mouldic, generally decreases with depth, but will be significantly enhanced at the transition zone between fresh and saline groundwaters. Mixing corrosion, which may or may not be accompanied by dolomitisation, will be particularly active at the margins of the island, where heterogeneous cavernous porosity is likely to develop. In the zone of saline groundwater, diagenetic evolution will continue, driven by large-scale circulation systems, and may result in precipitation of porosity occluding cements, or enhancement of cavernous and fracture porosity by dissolution and dolomitisation. Saline groundwater circulation will be particularly active near the steep margins of isolated carbonate build-ups, associated with vertically extensive bank-marginal fracture systems.

The best reservoirs have:
1. Prolonged and/or repeated development sequences
2. Low surface relief
3. High platform relief and steep margins
4. Large sealevel fluctuations
5. Seasonal climate with high effective rainfall
6. Unstable mineralogies
7. Dolomitisation
CHAPTER 2

PALAEOKARST: TYPES, RECOGNITION, CONTROLS AND ASSOCIATIONS

by V.P. Wright

2.1 INTRODUCTION

While karstic processes have traditionally been regarded as being related to near-surface meteoric (rain-sourced) waters, it is becoming increasingly clear that significant dissolution of carbonates can also take place in other settings, related to other fluids besides simply meteoric-sourced groundwaters. Figure 2.1 shows the major sites of dissolution diagenesis. Each is associated with its own fluid type and dissolution processes, but considerable overlap can occur between each. Meteoric and the burial karst facies are part of the continental hydrologic domain discussed in section 1.2.1. The chemical processes involved have been reviewed in section 1.1, but the distribution and styles of each type are the aspects relevant for hydrocarbon exploration and these will be the themes covered below.

2.1.1 Meteoric System

The most familiar karst products are those associated with meteoric water and most palaeokarsts are attributable to such origins. Surface and subsurface karst features develop: surface features vary from millimetre-scale dissolution 'scars' to large tower karsts, and fossil examples of many of these are known. Subsurface karst refers to cave systems and other conduits.

Most meteoric palaeokarsts have originated where shallow-marine limestones have become subaerially exposed by a fall in relative sea-level. The carbonate body is thus open to the atmosphere at its surface, forming an unconfined reservoir. Confined reservoirs are where the aquifer is confined by some overlying impermeable or poorly permeable layer.

A series of hydrologic zones can be defined (Fig 2.2) forming a meteoric-karst facies profile. The vadose zone is the unsaturated zone above the water-table through which water drains. The soil zone is underlain by the subcutaneous (or epikarst) zone which is highly weathered. Beneath is a less weathered free-draining zone. The water-table (or piezometric surface) is the surface where hydrostatic and atmospheric pressures are equal. It is mobile and a zone of intermittent saturation exists. Below is the phreatic zone, where all pores are saturated with water and within which broad sub-zones may be recognised related to the flow activities. Caves can develop in any of these zones, a cave being defined as any solution conduit larger than 5-15 mm (Ford 1988). The hydrologic characteristics of such channels are very similar despite their size variations.

The nature and extent of such cave systems has been extensively documented in many published works (e.g. Ford, 1988; Ford and Williams, 1989), and differences occur between those developed in vadose, unconfined phreatic and confined phreatic systems. The enormous amount of information on such cave system styles and on their porosity and permeability characteristics, available from studies of karst systems, appears at first ideal to apply to palaeokarsts. Indeed the sophisticated hydrological models discussed in this chapter might seem directly applicable to reservoir modelling. However, there are many problems particularly because many palaeokarsts have had complex, polyphase histories. It might be instructive to consider the stages of development that an uplifted carbonate terrain would undergo during prolonged karstification to illustrate this point.

The newly emerged sediments would both be unconsolidated and contain a variety of carbonate forms of different solubilities (low- and high-magnesian calcite, aragonite and possibly dolomite). The exposed surface would undergo dissolution, the degree of which would ultimately depend on the climate (see below). Initially, meteoric water would
Fig. 2.1 Occurrence of extensive carbonate dissolution diagenesis and main processes and agents responsible for the dissolution.
pass through bare sediment, or a thin soil cover and flow would be intergranular and intragranular (diffuse flow). With time cementation would take place, especially in the phreatic zone, and flow will become 'channelised' in conduits (conduit flow of White, 1969). The product of synchronous lithification and karstification is referred to as syngenetic karst (Jennings, 1971). Thus, primary porosity is reduced while secondary porosity increases. Depending on the climate and length of exposure, a wide range of porosities can result (Fig. 1.32).

While these processes are taking place, the carbonates are also becoming progressively stabilised to a low-magnesian calcite mineralogy (James and Choquette, 1984). Even with complete stabilization high primary porosities remain (Schmoker and Hester, 1986); only 15-20% of the original porosity being lost. Thus, for extensive cementation to have occurred there has had to have been either net loss of sediment by dissolution or by the addition of material from elsewhere. As conduit flow develops, caves result and the high flow rates in the shallow phreatic zone creates greater local dissolution. Such high porosity zones have even been recognised in palaeokarst reservoirs (Craig, 1988). It would clearly be of importance to be able to predict the thickness of zones of potential dissolution and karst porosity formation in ancient sequences. Recently Budd and Vacher (1991) have offered a predictive model for phreatic lenses beneath carbonate palaeo-islands, based on the Dupuis-Ghyben-Herzberg relationship (see section 1.2.2). The thickness of the lens (H) will be a function of the islands width (a), hydraulic conductivity (K) and the amount of recharge (R). The ratio of R/K in Holocene and Pleistocene sediments and rocks is in the range 10^{-4} to 10^{-6}. Thus ideally H should be 1-3% of the magnitude of the island width (a), but values of 0.2-2% are more typical with 1% being representative. Budd and Vacher use this approach on two reservoir intervals, the Smackover (Upper Jurassic) Oaks Field of Louisiana, and the Late Permian Yates Field of Texas (see Craig, 1988) which has a palaeocave system. A similar, though much cruder approach was employed by Watts and Wright (1978) on the Cretaceous (Edwards) reefs of Texas.

The flow rates through the conduit systems are much greater than through the host rock where diffuse flow operates. In the former, the rate is both rapid and turbulent, enhancing dissolution, while in the latter it is much slower and is laminar (White, 1977). This is a property which explains the complex behaviours of some palaeokarstic reservoirs where both cave and matrix porosity occurs. (Matrix is used here to refer to rock/sediment between the main conduits.)

The development of caves can be very rapid if a large source of water is available from outside the immediate catchment area (allogenic input), such as a nearby land area. An understanding of the regional paleogeography could be critical in assessing the potential for palaeokarst development in any sequence. If areas of localised autogenic recharge occur, such as dolines, flow rates and dissolution rates will be higher near them (Williams, 1985). Such features have been detected using 3D seismic (Brown, 1985), but the relatively small "catchments" of dolines are such that the actual amount of increased dissolution associated with them will be low compared with that of allogenic recharge areas. Palaeo-dolines associated with collapse features may also have enhanced porosities reflecting brecciated zones where collapse has occurred.

Flow, as stated above, is greatest in the upper phreatic zone and the less active zones may become stagnant. The upper phreatic zone is also oxic, although microbial activity can reduce this and the stagnant areas are typically sub-oxic to anoxic. This trend is also seen laterally when unconfined aquifers become confined down dip (Edmunds and Walton, 1983) (Fig. 1.7). This trend towards reducing conditions has commonly been found in studies of palaeoaquifer systems using cathodoluminescence behaviour as a guide to the redox conditions (Choquette and James, 1988).

Recognizing a palaeokarst in the geological record as of meteoric origin will depend on being able to relate its distribution to a particular exposure surface or unconformity. If the
Fig. 2.2 Hydrological zones in karst. Note contrast between the irregular "bare" and smoother, soil covered karst surfaces. Dissolution porosity formation is greatest in the more active upper part of the phreatic zone and also where mixing corrosion occurs.
palaeokarst is a single-phase system, i.e. it is not the product of more than one overlapping hydrological system, it may be possible to crudely predict aspects of the dissolution porosity zones. Cave porosity will be highest near the paleo-watertable and will decrease with depth. Zones of preferential higher porosity and permeability may be present near large autogenic drainage recharges, such as dolines, and may also be higher near allogenic recharge zones. However, not only does the system evolve from diffuse flow to conduit flow (with more complex porosity - permeability characteristics), but as base-levels change (due to tectonic, eustatic or even to geomorphic changes), karst zones will be overprinted.

2.1.2 Marine Mixing System

In the last few years, the importance of the marine-meteoric mixing zone has become widely appreciated. It is not only in such 'brackish' mixing zones that extensive dolomitization occurs (see review in Tucker & Wright, 1990), but it is a zone where very high levels of dissolution can also occur.

The main process creating karstic porosity is mixing corrosion (see section 1.1.3). Besides mixing corrosion other processes can operate and the microbial oxidation of organic matter, producing CO₂, is also an important process leading to undersaturation in marine-meteoric mixing zones (sections 1.1.2). An additional process in the mixing zones may be dissolution related to bacterial reduction of sulphate, present in the seawater.

Mixing zone waters can be especially aggressive and result in spectacular degrees of corrosion particularly in un-stabilised carbonate sediments. The result is a distinctive 'Swiss-cheese' style of dissolution illustrated by Back et al. (1986) and Smart et al. 1988a). Extensive cave development can also take place (Vernon, 1969; Leve, 1984; Mylroie, 1988; Smart et al., 1988a) and subsequent collapse can even control coastline shapes and sediment distributions (Hanshaw and Back, 1980; Back et al., 1984; Hine et al., 1988). The mixing zone is also one of dolomitization and highly vuggy and porous Quaternary dolomites, reflecting such settings, have been described (Ward and Halley, 1985).

Whereas mixing corrosion can take place in other settings besides marine-meteoric ones, it is these which are best documented and arguably of most relevance to hydrocarbon exploration. After all, most limestones were deposited in shallow water and were especially prone to subaerial exposure as a result of sea-level changes. This point has been dramatically shown in the study of Matthews and Frohlich (1987), who have computer-modelled shelf margin diagenesis. As their work shows, the opportunities for mixing-zone alteration of shallow-water limestones, using Quaternary sea-level changes as a guide, is considerable. It is likely that more examples of ancient mixing-corrosion zones will be discovered.

In general such zones will have developed in the coastal areas around carbonate terrains which were sub-aerially exposed and possessed a meteoric lens. Undersaturation can occur at a wide range of seawater-meteoric water mixtures (Back et al., 1986), even as high as 90% seawater for calcite (Stoessell et al., 1989). These coastal mixing zones will occur around the edges of carbonate shelves and carbonate ramps and also around and under isolated carbonate build-ups such as isolated platforms and atolls (see later section). Sandford and Konikow (1989), using models, have discussed the positions of mixing corrosion zones in relation to groundwater zones. They note that calcite dissolution and porosity formation occur primarily on the 'fresh' side of the mixing zone and in two distinct areas: one near the base, and one near the top where flow velocities are highest. The major control on such dissolution is the fresh-water flux, whereby much greater dissolution is associated with carbonates bordering large land areas than with small islands. Their models also show how rapidly such porosity could form, even over geologically short periods.

Mixing zones can also develop well offshore, where, if the hydrostatic head is sufficient, freshwater springs can emerge at the sea floor kilometres or
even tens of kilometres offshore (Johnson, 1983; Chafetz et al., 1988). Descriptions of mixing-zone palaeokarst are still very few in number. Mazzullo et al. (1990) have described extensive small scale dissolution porosity from the Upper Triassic Steinplatte and basal Liassic Hirlatz carbonates of the Northern Calcareous Alps of Austria. The cavity systems are highly complex in form, 0.3 - 0.4 m and larger in size, from vertically disposed to anastomosing networks with horizontal and subhorizontal orientations. These cavities are widespread through the limestones in the platform facies but not in fore-reef or hemi-pelagic carbonates. The cavities are filled by a range of replaced marine cements and internal sediments, the former occurring over a vertical thickness of 180 m.

Neptunian fractures, filled by fossiliferous Liassic sediments locally cross-cut the cavity systems. It seems likely that this was produced, at least in part, by mixing dissolution. The occurrence of vadose fabrics indicates that some developed in the upper meteoric zone.

The importance of mixing zone dissolution in hydrocarbon reservoirs has been noted by Craig (1988) from the Permian Yates Field of West Texas and Bouvier et al. (1990) from the Amposta Marino Field, offshore northeast Spain. It is still rather 'early days' in the study of palaeokarst facies to define the characteristics of a palaeo-mixing zone reservoir, but some general points may be relevant. The zones will probably be highly porous with irregular vuggy, earthy porosities associated with dolomite in some cases. The, probably, penecontemporaneous dissolution of calcite (or aragonite) and precipitation of dolomite will create an unusually complex porosity distribution. The individual zones will probably be relatively thin up to perhaps ten metres, but circumstances could have developed allowing thicker ones to have formed. In shelf or ramp settings they will probably thicken seaward, while on low relief isolated build-ups they may be very extensive laterally (e.g. Andros Island, see chapter 2). The extensive breccia-fracture reservoirs of the Ordovician Ellenberger Group of west Texas may be a candidate for such a platform-wide mixing zone system (Kerans, 1988). Extensive mixing-zone palaeokarst also occurs in the Devonian platform limestones of the Great Slave Lake, Canada. These host the Pine Point ore deposits (Ford, 1990). However, considering the importance of mixing-zone processes during the present global high stand of sea-level, the apparent rarity of mixing-zone palaeokarst is striking.

2.1.3 Deep Burial Fluid Karstification

Dissolution can occur in deeper burial settings, but will not create large-scale porosity (caves) because of the high lithostatic pressures. Its effects are commonly seen in many diagenetic sequences where 'late' burial secondary porosities develop, often associated with hydrocarbons and mineralisation.

Several processes operate to cause the undersaturation of calcium carbonate, and these operate over a wide depth range. They are related to burial processes, especially the maturation of organic matter, and the migration of oil field gases and fluids and other thermal fluids. However, extensive dissolution porosity may develop in shallower settings where cooling or mixing with oxygenated, cooler meteoric waters takes place (see sections 1.1.4 and 1.1.5). The two main agents of dissolution are CO2 and H2S. The former is typically derived from the decarboxylation of organic matter during hydrocarbon maturation. Thus, extensive dissolution porosity can be created in association with migrating oil (Druckman and Moore, 1985).

Whatever the exact source of the CO2, thermal waters enriched in it can migrate into shallower burial settings and create significant dissolution porosity as the waters cool and mix with meteoric water. While mixing corrosion may also occur, it is the cooling of the waters which is probably the major cause of dissolution. The solubility of calcite is greater at low temperatures and as the fluids migrate they cool and cause dissolution (Wood, 1986). This possible role of CO2 - enriched thermal fluids has been discussed by Ford and Williams (1989) and spectacular examples occur in
the Jewel and Wind caves of South Dakota (Bakalowicz et al., 1987) (Fig. 1.16) and in eastern Europe. Ford and Williams have suggested diagnostic features for thermal CO₂ systems such as the development of highly complex, multi-storey, maze-like cavern architecture. These systems lack typical 'meteoric' cave sediments but are associated in many cases with hydrothermal mineralization and extensive brecciation. As such many hydrothermal karsts are associated with Mississippi Valley-type deposits (Dzulynski and Sass-Gustkiewicz, 1989). This type of hydrothermal karst produces several distinctive styles of cavity, including narrow sheet-cavities in closely-spaced parallel sets. Such hydrothermal palaeokarsts are also associated with extensive burial dolomitization, of a distinctly different style to that associated with marine-meteoric mixing-zone processes and easily distinguished using standard petrological techniques.

Hydrogen sulphide (H₂S) is also an agent of dissolution when it reacts with water to form sulphuric acid. This occurs by mixing with shallow, oxygenated meteoric waters. The acid reacts with the limestone forming gypsum which can then later dissolve to form larger caves. Further details are given in chapter 1.1.5.

Burial fluid-related palaeokarsts may also form in association with unconformities, in positions where CO₂ or H₂S-related dissolution could take place as these fluids cooled and/or mixed with meteoric waters. The palaeokarstic zones are likely to be related to particular palaeo-fairway zones which could be predicted from a knowledge of the regional geology.

2.2 PALAEOKARST TYPES

Palaeokarst refers to karstic (dissolution-related) features formed in the past, related to an earlier hydrological system or landsurface (modified from Wright, 1982).

Karst systems have been intensively studied and the hydrological principles of their formation are well understood. Thus, in studying palaeokarsts it is possible to employ the knowledge of present-day systems to understand ancient ones. Indeed this is one of the few occasions when 'diagenesis' can be related to hydrology and hydrogeology, allowing some predictive element.

There are two main drawbacks encountered in studying palaeokarsts which frustrate attempts to apply karst-science to interpret them. Firstly, many palaeokarsts are encountered in the subsurface, particularly during hydrocarbon exploration, and interpreting palaeokarst style from limited seismic or bore-hole data is very difficult. At outcrop, the task can be easier but the second major problem is that many major palaeokarsts are polyphase and are difficult to interpret for this reason. Major palaeokarsts are either associated with unconformities or high amplitude sea-level falls. In both cases, long exposure periods and complex base-level changes result in multiple phases of development making formulating simple palaeohydrological models difficult.

The majority of palaeokarsts, however, are of minor development (but can create significant porosity). They are associated with small-scale sea-level changes (e.g. fourth- and fifth-order changes) and should be regarded as an integral part of the diagenetic spectrum for shallow-marine limestones. After all, most limestones were deposited in very shallow water (less than 10 m) and were as such, vulnerable to even small-scale changes in sea-level and related meteoric diagenesis.

At the simplest level two simple divisions of palaeokarst can be made, as in the case of karst, between surface palaeokarst (formed at an ancient landsurface) and subsurface palaeokarst (formed beneath the ancient landsurface). A special case exists, of subjacent karst which is a form of subsurface dissolution which can mimic surface dissolution.

Broadly speaking three types of palaeokarst can be recognised (Wright, 1982): relict, buried and exhumed (see also Bosák et al. 1990, p.25-32 for a review of terminology).

(1) Relict palaeokarst represents karst which developed under different
climatic conditions from which it is now found. It relates mainly to Cenozoic karst and is of no significance in an exploration context.

(2) *Buried palaeokarst* is ancient karst overlain by younger sediments. Karstification took place prior to burial (c.f. interstratal - karst).

(3) *Exhumed palaeokarst* refers to palaeokarst which although once buried, is now being exhumed and exposed to further karstification. This is also termed fossil karst.

The most important category for hydrocarbon exploration is buried subsurface palaeokarst, but surface palaeokarsts can, when detected, provide subtle clues to palaeoclimates and porosity evolution (see below).

*Interstratal karst* traditionally refers to the development of dissolution features on a soluble host rock at its upper contact with a cover of permeable (and usually siliciclastic) material. The term has been used synonymously with subjacent or subjacent karst. Slight confusion over terminology can arise because the term interstratal has been used in, for example, sandstone diagenetic studies to refer to any subsurface dissolution. It is proposed here that its use be modified and used to refer to dissolution of rocks along bedding planes or unconformities, in the subsurface. The dissolution of evaporite beds is the simplest case.

*Subjacent karst* should refer to karst features developed at the contact between overlying permeable, typically siliciclastic deposits and underlying lithologies. For subjacent karst, the simplest case to imagine is where porous sediment (e.g., sands or sandstones) overlay limestones. Water may move down through the upper unit (depending on the water table level) and corrode the top of the underlying limestone (Fig. 2.3). In some cases this leads to the formation of cavities at the contact, followed by collapse of the overlying unit to form karstic (doline) features on the non-carbonate unit (Fig. 2.3). The best-known example is the Millstone Grit (Namurian) - Carboniferous Limestone (Dinantian) contact in South Wales. As a result of collapse at depth (over 40 m in places), the surface of the Namurian pebbly quartzites is pock-marked with large "doline" structures. This dissolution of the Carboniferous Limestone is believed to have been controlled by the positions of earlier (pre-Namurian) paleokarstic topography, (legacy karst, see below). This example (see Wright 1986 for details and references) has wider implications for the Mississippian-Pennsylvanian contact in many areas (U.S., Europe), is an unconformity between carbonates and overlying siliciclastics.

The main problem for exploration is that interstratal and subjacent karstification can produce features resembling palaeokarstic surfaces. For example, a shallow-marine limestone overlain by sandstones, may later undergo interstratal dissolution, either early or late in its history, to form a karst horizon at its upper contact which could be misinterpreted as a true surface palaeokarst. Virtually nothing is known of the styles of karst associated with this type of dissolution. Criteria for differentiating subjacent from buried surface palaeokarsts have been discussed by Wright (1982). Extensive subjacent dissolution will result in the collapse of overlying beds which may be a signature for subsurface recognition.

To summarise, subjacent and interstratal karst are special cases of subsurface karst. Subjacent karst refers to dissolution at a lithological discontinuity between overlying permeable strata and underlying soluble lithologies. Interstratal karst refers to dissolution with a buried soluble unit, such as buried evaporites, typically along bedding planes or an unconformity.

2.3 POLYPHASE KARSTIFICATION AND PALAEOKARSTS

Many carbonate sequences exhibit multiple phases of karstification and an appreciation of this should prove useful in interpreting complex palaeokarst systems. Such phases may be multiple, but result in discrete palaeokarst features, whereas others overprint
Fig. 2.3  A. Formation of subjacent karst. Dissolution occurs beneath impermeable cover.
B. Example on Carboniferous of South Wales (see text).

Fig. 2.4  Polyphase stacked palaeokarsts on a carbonate platform. Similar subsurface karst systems probably occur on the Bahamian platforms related to various Pleistocene low stand mixing zones. The preservation of palaeokarstic surfaces depends on the configuration of the platform and on its sea-level flooding history. On silled platforms the surfaces have generally been strongly modified or destroyed around the margins as a result of physical and biological erosion. The late Mississippian platform sequences of Britain contain large numbers of stacked palaeokarstic surfaces. See text.
earlier phases.

The simplest case is where multiple changes in sea-level result in stacked karst systems. Under Andros Island, (Fig. 2.4), on the Great Bahama Bank, there is a major cavern system related to the last low-stand mixing zone (Smart et al., 1988a) and other systems may exist below this one related to other mixing zones.

Stacked surface palaeokarsts have been documented from the early Carboniferous of Britain. Walkden and Davies (1983) have described a large number of palaeokarst surfaces and associated fluvial intercalations from the Asbian and Brigantian (late Dinantian) of North Wales. The ability to recognise such palaeokarstic surfaces in outcrop or subcrop is clearly important, but the preservation potentials of such surfaces is a critical factor to consider. Rasmussen and Neumann (1988) have discussed this aspect in relation to exposure surfaces in the Pleistocene of the Bight of Abaco, Bahamas. They found that the final record of exposure within the platform deposits was controlled by variety of factors, such as topography, sea-level flooding history and the position of the surface within the flooded platform. Whether the platform is silled or open is critical: in the case of the former, the exposure surfaces are best preserved in the deeper, interior areas of the platform lagoon, while they are modified or destroyed by bioerosion around the elevated platform margins. In more open platform settings, both physical erosion and bioerosion reduces the preservation potential of the surface during sea level rise. On ramps such surfaces may be preserved if buried by low energy back-barrier deposit during transgressions associated with barrier shorelines (Riding and Wright 1981).

It is common for later phases of karstification to overprint earlier ones. One case was noted earlier in the Mississippian of South Wales, where palaeokarst developed at the top Dinantian unconformity may have influenced later (Quaternary or Tertiary) subjacent karstification. The term legacy karst might be appropriate for such a situation (legacy karstification refers to dissolution occurring at the present or in the past whose distribution is controlled by an earlier (palaeo) karst system (Fig. 2.5A). In such a case, the overlying units will show signs of collapse and not simply passive infilling on a non-active dissolution surface or zone.

Another simple case occurs where a landscape is undergoing erosion and earlier phreatic sub-surface karst is emplaced in the vadose zone (Fig. 2.5B). Such an example has been described from Precambrian palaeokarst from the North West Territory of Canada by Kerans and Donaldson (1988). In this case a broad carbonate platform of Middle Proterozoic age was subaerially exposed. Initially karstification took place in clearly-defined vadose and phreatic settings, the latter associated with caves. Subsequently, as a result of falling base-level and erosion, the phreatic caves were modified in the vadose zone and had a variety of sediments deposited in them by free-flowing streams. It may be more realistic to envisage the reverse situation for many ancient carbonate sequences, where phreatic, subsurface karst may overprint earlier vadose karst. In the Amposta Marino field, offshore northeast Spain, the karstified lower Cretaceous Montsia Limestone is sealed by Miocene shales at a "buried-hill" trap. The main high porosity zone is strikingly continuous and subhorizontal, quite unlike a simple fracture controlled phreatic karst system. Wigley et al. (1988) and Bouvier et al. (1990) have interpreted this as a marine-mixing corrosion zone, overprinting the original phreatic system as a result of a net sea-level rise (Fig. 2.6). Again this serves to illustrate the common occurrence of polyphase karstification.

The situation can be even more complex in carbonate sequences exposed either to prolonged, continuous exposure or to complex burial histories. Two examples have been documented which illustrate long-term polyphase karstification: the Turonian limestones of Israel, and the Carboniferous Limestone of South Wales.

2.3.1 Turonian to Neogene Palaeokarst of Israel

The extensive karstic systems found today in the Cretaceous carbonates in central and northern Israel have been
developing for over 80 Ma (since late Turonian times) (Buchbinder et al., 1983). In late Turonian times exposure resulted in the formation of small dolines (20-100 m in diameter) within the upper Turonian limestones. During the Senonian, two subsequent phases occurred, in Coniacian and Santonian times. These resulted in the development of larger dolines, filled by contemporaneous sediment. Another phase of pre-Middle Eocene doline development occurred followed by an extensive phase of Neogene karstification. This was characterised by deep subsurface dissolution.

Not all phases occur in all areas; this partial absence is a common feature of regional polyphase karstification. This reflects areal variations in emergence. While much of the study material described by Buchbinder et al. was seen at outcrop, deep boreholes have encountered Turonian-hosted palaeokarst in the subsurface of the Israeli coastal plain.

2.3.2 Carboniferous Limestone of South Wales

The Dinantian (Mississippian) limestones of South Wales also illustrate the problems of multiple exposure to karstic system development. The limestones, over 1500 m thick in places, were deposited during the ramp to shelf evolution, followed by foreland basin development. The sequence contains a number of palaeokarsts (Wright 1986) which are strongly areally differentiated. Differentiation of the province during foreland basin evolution also occurred, with the emergence of the present northern outcrop areas as a foreland bulge in late Viséan times. In contrast, the present southern outcrops subsided very rapidly at the end of the Dinantian and were only unroofed in the Mesozoic (see later section on foreland basin palaeokarst).

Both the northern and southern outcrops underwent numerous phases of exposure during Mississippian times. The northern area was also affected by local phases of uplift related to active strike-slip faults and has slightly more exposure surfaces than the southern sequences (Figs. 2.7 and 2.8). The uplift of the northern areas as a foreland bulge at the end of Mississippian times led to extensive erosion. Here, the top Mississippian unconformity is a palaeokarstic surface, later modified by interstratal dissolution (Fig. 2.3), acting along the limestone unconformity surface, apparently following and enhancing relief on that surface (legacy karst). As a result of subsurface cavern collapse at the top of the limestone, the overlying Namurian (Pennsylvanian) quartzites collapsed locally and presently exhibit a doline-like morphology at the surface. However, an alternative explanation has been offered by Battiau-Queney (1986) who interprets the quartzite doline terrain as reflecting tropical silicate weathering during the early Tertiary. She also recognises tropical karst features in the local limestones. Thus a Tertiary phase of karsting may also have occurred (Fig. 2.7A).

In the southern outcrops (Fig. 2.7C) the limestones were deeply buried in a developing fore-deep and were apparently not exposed at the end of the Mississippian. They were not exhumed until the Triassic or Jurassic and cave fills of these ages are known. A probably Cretaceous cover was not removed until after early Tertiary tropical weathering which is not evident in the region, only along the northern outcrops (Fig. 2.7).

2.4 SUBSURFACE RECOGNITION

The various techniques which have been used to detect palaeokarst zones will be briefly reviewed, but few studies have been published. Often such zones are indicated by a loss of circulation or the dropping of a drill bit when a large cavity is breached. Palaeokarsts have been detected by wire-line techniques. For example, gamma-ray logs may show progressive up-hole peaks reflecting terra-rossa deposits (clay-filled karstic cavities). Sonic logs have also shown up-hole gradients related to the degree of fissuring and brecciation (Del Olmo & Esteban, 1983). In both cases, vertical trends are useful for recognition. Dip-
Fig. 2.5 Polyphase karst.
A. Buried palaeokarst influences later subjacent dissolution (e.g. Carboniferous Limestone of South Wales; see text.)
B. Phreatic karst later overprinted by vadose processes (see text for details.)

STAGE I — Compartmentalised meteoric phreatic system

STAGE II — Mixing zone overprint (improved communication between porosity network)

Fig. 2.6 Schematic diagram of polyphase development of tilt-block palaeokarst. Based on Amposta-Marino Field, offshore north east Spain (see text).
Polyphase karstification of the Carboniferous Limestone (Dinantian Mississippian) of South Wales. A. Karst history of northern outcrops. B. Locality map. C. Karst history of southern outcrops. See Fig. 2.6. (Modified from Wright, 1986.)
Fig. 2.8 Contrasting subsidence and karst histories of the Carboniferous Limestone in South Wales. A. Northern outcrops. B. Southern outcrops. Numbers refer to those in Fig. 2.5 A & B.
meter data can also indicate drape over palaeo-(karstic)-topography (Vandenberghe et al., 1986). Density log, caliper and bulk-density porosity logs can also show karstic porosity, down to small caverns with heights of 0.3m (Craig, 1988).

There are several methods whereby palaeokarst has been detected by seismic techniques (Fontaine et al., 1987):

1. Erosional truncation of reflectors at an unconformity will suggest the possibility of palaeokarst on limestone sequences (Del Olmo and Esteban, 1983).
2. Topography on limestones, unconformities and hence the possibility of palaeokarsts, can be revealed by onlap of reflectors onto the unconformity (Fig. 2.9).
3. Collapse zones may be revealed by abundant irregularities affecting the seismic marker (Jenyon 1984, 1986).
4. 3-D seismics have been used to identify karstic depressions (Brown, 1985 and even palaeo-mixing corrosion zones (Wigley et al. 1988); Bouvier et al., 1990).
5. Amplitude analyses (offset-seismic) has proved a useful tool (Vandenbergh et al., 1983, 1986). In simple terms, it is a technique which reflects the density of the limestone, and is useful for detecting highly fissured zones. Vandenberghe and co-workers tested this technique on the major sub-Namurian palaeokarst in northern Belgium. It has also been used in the off-shore palaeokarst reservoirs of N.E. Spain (Fontaine et al., 1987). Bouvier et al. (1990) have used amplitude differences to map palaeokarstic porosity distribution on the Amposta Marino oil field offshore Sapin. Areas of high amplitudes corresponded to highly cavernous zones around the buried 'hill' (Fig. 2.6) where cavern collapse had occurred, with porosity/permeability later enhanced by zones of mixing corrosion dissolution. Areas of low amplitude occur around the crest of the field where less significant fissure porosity occurs.

Cores have particular limitations for evaluating palaeokarstic features because of the often large size (relative to core width) and irregular orientations of karstic features. Such porosity would be classified as vuggy, channel and cavern porosity: but many "fractured" reservoirs are also probably palaeokarstic in origin. The Ordovician Ellenberger Group reservoirs of west Texas are a striking example of such a system (Kerans, 1988). This extensive reservoir interval had previously been interpreted as a tectonically fractured dolomite, whereas detailed subsurface studies showed that the fracturing was clearly related to an overlying unconformity. The fracture-breccia porosities appears to have resulted from the collapse of large scale cavern systems developed by an extensive mixing zone under a carbonate platform (Kerans, 1988). Clues to the palaeokarstic origin of irregular porosity would include large non-fabric selective solutional pores, which may be open or partially filled with internal sediments or cements. Speleothems (commonly coarsely crystalline calcite cements) are an obvious feature but are not always found associated with palaeokarsts. Argillaceous materials, often red, are a common feature, and are generally referred to as "terra-rossa". Such material occurs near the top of a karstic profile but may be washed down to considerable depths below an unconformity. It may have been derived from dissolution of limestone as residual insolubles, or it may be aeolian in origin.

Other exposure related features may provide clues to the nature of the irregular porosity zone, such as paleosols including calcretes. Esteban and Klappa (1983) have reviewed many of these criteria.

2.5 CONTROLLING FACTORS ON KARST DEVELOPMENT

2.5.1 Palaeokarst and Palaeoclimates
Fig. 2.9  Simplified seismic section through the Bresse Basin (France) showing Eocene-Oligocene onlap on a Jurassic palaeokarstic high. Well three encountered unkarsted Mesozoic limestone. The palaeokarst is limited to the high. Based on Fontaine et al., 1987.

Fig. 2.10  Contrasting styles of karstification seen in the early Carboniferous oolites of South Wales (based on Tucker & Wright 1990).
While the style of meteoric diagenesis and karstification a carbonate unit undergoes will, in part, be controlled by whether the material is unconsolidated (diffuse flow) or lithified, bedded and fractured (free or conduit flow), the main controls will be climate, composition, relief (in relation to sea level changes) and time.

In this section the influence of climate is discussed both with respect to karst morphology and also in relation to general porosity evolution during meteoric diagenesis. Climate is important not only in its influence on the availability of meteoric water, and its flux, but also for temperature, and related factors such as soil cover and vegetation (Ford and Williams, 1989) (chapter 1).

The effects of climate on the rates of karst denudation have been discussed in detail by White (1984) (see Chapter 1) and the rates are highest in regions of high effective rainfall (precipitation minus evapotranspiration) and high CO₂ levels in the soil.

Under arid conditions, low rainfall will result in relatively slow rates of meteoric alteration (dissolution, cementation and mineralogical stabilisation), and little water will reach the aquifer through the vadose zone. Secondary carbonates will accumulate slowly in the upper part of the vadose zone, typically as calcite, and the water table will be deep. Low rates of groundwater movement will result in little phreatic cementation. Under wetter conditions net dissolution may occur with extensive mouldic and karstic porosity formation, with little secondary carbonate accumulation in the vadose zone (no calcrite). However, as a consequence of high rates of dissolution and a high flux, extensive cementation in the upper phreatic zone. These two scenarios are end-members and most sub-tropical carbonates undergo meteoric diagenesis in an intermediate setting.

The influence of climate on the whole evolution of porosity in subaerially-exposed carbonate sediments has been documented from several Quaternary and pre-Quaternary sequences. A comparison of Pleistocene and Holocene carbonate aeolianites of north-east Yucatan is particularly instructive in this sense (Ward, 1978; McKee and Ward, 1983). In the region, in late Pleistocene times the aeolianites were subaerially-exposed under an arid-climate. Whereas unconsolidated carbonate sands underwent little secondary porosity formation and little cementation, they exhibit pedogenic (calcrete) features such as finely crystalline calcite cements, rhizocreations, needle-fibre calcite and calcrite crusts. In the younger Holocene aeolianites, secondary porosity is more extensive and coarse calcite cements are common, including even some of the youngest deposits being locally totally cemented. However, rhizocreations, needle-fibre calcite and calcrite crusts are absent. The present day climate is humid. During the late Pleistocene the more arid climate resulted in a lower flux of meteoric waters and less dissolution and less cementation. Similar climatic controls on porosity evolution have been documented from the Pleistocene of Mallorca (Calvet et al., 1980) and Barbados (Harrison, 1975).

An ancient analogue for the climatic controls on subaerial diagenesis has been recorded from the Mississippian oolitic grainstones and peritidal limestones of South Wales. These examples suggest a possible means of predicting the porosity evolution of subaerially exposed carbonates based on assessing the palaeoclimate during exposure from the nature of the exposure surface itself and this should be possible even from core material. Within the various oolitic sand bodies comprising the Mississippian succession in South Wales, two styles of meteoric diagenesis can be recognised:– semi-arid and humid (Hird and Tucker, 1988; Wright, 1988). The oolites exposed under semi-arid conditions exhibit minor karstic surfaces of the mammillated type (Fig. 2.10) which are veneered by thin calcrite crusts with abundant rhizocreations and needle-fibre calcite. The associated oolitic grainstones show virtually no evidence of significant pre-compaction(pre-burial) cementation. A younger oolite exhibits exhibits, locally, total early phreatic cementation. The exposure surface is characterised by an intense zone of solution piping (a form of kavernossed karren, see below), which is a style more typical of humid
climates. Calcrete crusts, rhizocretions and needle fibre calcite are absent.

The diagenesis of the weakly karsted oolite can be compared with the late Pleistocene of Yucatan, while that of the other resembles the present day situation. These differences are shown diagrammatically in Fig. 2.10. Other formations occur in the succession also referable to semi-arid or humid phases (Wright, 1988).

2.5.2 Secondary Porosity and Temporal Changes in Carbonate Mineralogy

Diagenesis is driven by the instabilities of carbonate sediments and rocks. During burial diagenesis the driving forces are enhanced temperatures and pressures whereas many of the important changes during diagenesis occur very early on in the burial history because of the mineralogical instability of carbonate sediments in near-surface meteoric conditions. This instability is the driving force for much of "early diagenesis" and is especially important as it influences the degree of dissolution diagenesis (see Chapter 1). Marine carbonate sediments in today's oceans consist of calcite and aragonite. The calcite is in two forms: low magnesium (or magnesian) calcite (LMC) (<4 mole % Mg) and high magnesium calcite (HMC) (>4 mole % Mg). Both aragonite and HMC are replaced during diagenesis by LMC. Aragonite is replaced, usually, via a dissolution void stage (HMC is replaced by very fine scale dissolution/precipitation which does not result in porosity formation). Sediments rich in aragonite (e.g. some oolites, most post-Palaeozoic shallow water bioclastic limestones) will undergo considerable dissolution, and mouldic porosity will form, commonly later occluded by cement. Calcitic material can, of course, also undergo dissolution.

Clearly, mouldic porosity will be greater in originally aragonite-dominated limestones than in non-aragonitic ones. Indeed, overall solution rates will be greater in aragonitic host rocks, and thus we can refer to a sediment's "diagenetic potential" in relation to its susceptibility to dissolution diagenesis.

The diagenetic potential of marine limestones has varied through geological time because of two factors:

(1) As a consequence of biological evolution, bioclastic limestones have had different bulk compositions through time. For example, in broad terms, shallow-marine bioclastic and reefal limestones in the Palaeozoic were calcite dominated. Since early Mesozoic times, with the diversification of molluscs and the appearance of scleractinian corals (both aragonitic), the diagenetic potential of bioclastic limestones has increased.

(2) The mineralogy of non-skeletal marine carbonates is controlled by a number of factors, especially Mg/Ca and PCO₂. These have varied through geologic time with the result that the mineralogy of non-skeletal precipitates (e.g. ooids, lime muds and cements) has also varied. These variations, of calcite or aragonitic phases, have been plotted to produce what is known as the Sandberg curve (Sandberg, 1983) (Figs. 2.11 and 2.12). For example, during the Ordovician to the Mississippian, ooids were calcitic, with a low diagenetic potential for mouldic porosity. Mid Carboniferous to Triassic ooids, however were aragonitic and had a high potential for mouldic porosity formation. Local environmental effects can overprint this general trend and exceptions are known.

In addition, other factors have varied through geologic time, e.g. the PCO₂ of the atmosphere (Greenhouse phases; Fischer, 1981) (Fig. 2.12) and the degree of biological activity on the land surface (affecting soil cover). Both of these inter-related factors will have influenced paleokarst development. Even though broad speculations can be made about atmospheric composition changes through geologic time, and the biofunction of soils, a more striking relationship is emerging between carbonate mineralogy and sea-level changes (see below).
Fig. 2.11 Diagenetic potential of limestones through geologic time. The Sandberg curve indicates phases of different mineralogies of abiogenic carbonates. Aragonite phases have the greatest potential for secondary porosity development. However, long term changes in the skeletal composition of bioclastic material also needs to be considered. Lower Palaeozoic shallow marine limestones contain mainly calcitic bioclasts, while Mesozoic and Tertiary bioclasts were dominantly aragonitic.
Fig. 2.12 Upper diagram shows Sandberg curve for aragonite-calcite marine precipitates (see Fig. 2.9). The mid part shows phases of high PCO₂ ("greenhouse effect" = G) and low PCO₂ ("ice house phases" = I).
2.5.3 Sea Level Changes and Palaeokarsts

Relative sea-level falls are, of course, a prerequisite for the subaerial exposure of subaqueously deposited carbonates. In relation to this, two aspects require consideration. Firstly, the actual amount of fall, or base-level change, is critical in controlling the style of karstification. Secondly, there is a general relationship between global sea-level and carbonate mineralogy (Wilkinson et al., 1985; James and Choquette, 1984).

In the latter case it has been found that during major high stands calcite was the predominant marine precipitate, and aragonite during low stands. The nature of the relationship is believed to be controlled by geotectonics. High stands reflect periods of major plate movements, when hydrothermal weathering of basalts causes Mg depletion (lowering Mg/Ca and so promoting calcite, not aragonite, formation). In addition, high plate activity affects PCO₂, for it is believed that much of this is produced at subduction zones during metamorphism.

High PCO₂ and high sea-levels will create 'greenhouse global climates' and enhanced karstification (James & Choquette, 1984). It is possible to make the general statement that during high stands (e.g., Cambrian - early Carboniferous, and Jurassic - Cretaceous) rates of karstification were probably high (high PCO₂) and during the Jurassic - Cretaceous the overall diagenetic potential was higher because of abundant aragonitic biotas. During the low-stand phases (Permian - Jurassic and Tertiary to present day) more limestones may have been exposed but under lower PCO₂. However, the overall diagenetic potential was higher, because of the greater amounts of aragonite in the sediments.

Thus any attempt to predict the abundance of palaeokarst through time needs to consider a number of conflicting factors: sea-level, skeletal composition, non-skeletal carbonate composition, PCO₂. Ideally, a combination of low stand (hence widespread exposure), high aragonite skeletal composition, high aragonite non-skeletal composition and high PCO₂ would be required. A study of Fig. 2.12, taking the Vail et al. curve, reveals a phase of low stand through the late Triassic to early Cretaceous, corresponding with a greenhouse phase and with a high diagenetic potential because of the abundance of aragonite skeletal components. However, to date no compilation of palaeokarst abundance in the Phanerzoic is available to test this idea, although Mesozoic bauxites associated with palaeokarst are well documented. It remains to be shown if the late Triassic to early Cretaceous was a time of enhanced karstification.

The other influence on sea level is on the hydrological development of karst, controlled by the amplitude of the changes. This factor will be most clearly seen if we consider the effects of different orders of sea-level change (Fig. 2.13). During small, fourth or fifth order changes, the residence time of the carbonate sediments in the meteoric zone will be relatively short, perhaps only a few thousand years. Such small-scale falls will create little topographic relief and the flux of meteoric waters will probably be small. As a result relatively little dissolution and cementation will occur. Most flow will be of the diffuse-type, and the opportunities for karst development will be minor.

During progressively greater amplitude changes (third and even second order sea-level cycles) such effects will be much greater, reflecting longer residence times, increased relief and deeper flow, and, as flow passes to conduit type, larger scale karstic features including caverns, will develop. The relationships between major palaeokarsts and second/third order sea-level changes are discussed in Chapter 3.

Most palaeokarstic zones encountered in the geological record are those representing the higher frequency 'Milankovitch', orbitally-forced (fourth and fifth order) cycles and it is worth considering how changes in the magnitude of the cycles through the Phanerzoic may have controlled not only the styles of palaeokarsts, but also the architectures and compositions of the host carbonates (Fig. 2.14).

The precession, obliquity and eccentricity cycles are now known to be very important in controlling carbonate deposition in the geological record.
Fig. 2.13 Schematic relationship between orders of sea-level cycles (coastal onlap curves) and palaeokarst development. (Modified from Kerans, 1989).
Fig. 2.14  A. Fourth and fifth order sea-level curves during 'no-ice' phase. B. Depositional and early diagenetic styles seen in carbonate sequences deposited during high frequency sea-level changes without significant continental ice-buildups, e.g., Triassic (Ladinian) of northern Italy (see text).
These former two cycles have varied in length during the Phanerzoic as a result of the evolution of the earth/moon system but a more significant factor relating to the potential of all these to affect both carbonate deposition and karstification has been the changed degree of continental ice build-up, itself dependant on both global palaeogeography and climatic change. We contend here that the propensity for ice build-up is an important factor in controlling the types of palaeokarsts developed in carbonate build-ups.

During periods when these high frequency sea-level changes occur without significant ice build-ups, the amplitudes of the sea-level changes are very small, in the order of a few metres (Koerschner and Read, 1989). These small sea-level falls may occur at slower rates than ones driven by ice build-up, and are outstripped by regional subsidence. The later, small, relatively slow sea-level rises are outpaced by carbonate productivity. As a result, the carbonate platform is essentially one exhibiting 'keep-up' characteristics resulting in thick sequences of stacked, small-scale peritidal cycles. In such cases carbonate ramps rapidly evolve into aggradational shelves in the sense of Koerschner and Read (1989).

In summary, the carbonate build-up will consist, architecturally, of stacked, shallowing-up cycles representing precession cycles, perhaps arranged in penta-cycles representing the eccentricity changes (Goldhammer et al., 1987). Exposure events will be both short lived and the small sea-level falls will have created little relief on the platform. The effects of any eccentricity-related changes will be smothered by subsidence and productivity factors. Opportunities for karst development will be very minor but under suitably arid climates extensive sabkhas, evaporites and dolomites may form. The reservoir potential of such aggraded platforms will reflect the generally low permeabilities of the peritidal sequences (unless dolomitised) although the likelihood of stratigraphic trapping, associated with evaporites, may be high. Palaeokarstic plays may be unimportant, but intraplatformational basins may have developed, with their high source-trap potential, as well as peritidal/sabkha plays. Ancient platforms exhibiting this aggradational style include the Proterozoic Rocknest platform of Canada (Grotzinger, 1986), the Cambro-Ordovician of the eastern U.S. (Koerschner and Read, 1989) and the Triassic Italy (Goldhammer et al., 1987).

If major ice build-ups occur the facies and diagenetic style is changed radically to one, potentially, favouring major meteoric alteration and karst development. This situation is clearly exhibited in the late Quaternary record of the Great Bahama Banks (Fig. 2.15). As a consequence of ice build-ups the amplitude of both the falls and rises is large. Major falls, typically rapid, result in prolonged exposure and relatively deep meteoric circulation. As a consequence karst development occurs, including significant mixing-zone dissolution (Smart et al., 1988a). The rapid "ice-melt" sea-level rises cause initial platform drowning and typically only reef growth around the margins of the platform is capable of keeping-up with such rises, creating steep rimmed margins. The platform interior will contain predominantly subtidal deposits and patch reefs with minor peritidal facies (Beach and Ginsburg, 1980; Koerschner and Read, 1989). Under such conditions carbonate ramps will show less aggradation and may be dominated by grainstone wedges capped by exposure surfaces, onlapped by offshore ramp deposits.

Palaeokarstic hydrocarbon plays will be major targets in such carbonate build-ups. Architecturally the platforms may consist of subtidal, low-to-high energy, carbonates, and both meteoric and mixing-related porosity formation will be likely to be significant. Under more arid conditions, evaporites may form during sea-level down-draw. The steep, reeval platform margins may act to enhance circulation (and diagenesis) within the platform related to such factors as geothermal gradients (Whitaker and Smart, 1990) (Chapter 1).

The Bahama Banks might provide a present day analogue, and the Mississippian of Britain provides a very striking, ancient one. During late
Fig. 2.15 A. Example of ice-enhanced sea-level changes - Pleistocene sea-level fluctuations. B. Palaeokarst - platform architecture during periods of significant continental ice buildup and melt. Relatively rapid falls and rapid rises result in a sequence of predominantly subtidal (catch-up) deposits and major exposure surfaces (based on late Mississippian of Britain).
Mississippian times (Asbian and Brigantian stages) Britain (and also northern France, Belgium and southern Germany) was covered in a series of carbonate shelves and isolated platforms. As a consequence of asymmetric sea-level changes, with both a rapid sea-level rise and relatively rapid falls, the sequence throughout the area, is characterized by cycles of shallow, subtidal carbonates capped by prominent palaeokarstic surfaces (Walkden, 1974; Walkden and Walkden, 1990). Horbury (1989) has described the styles of Asbian cycles in detail and has argued for them reflecting a 100,000 year-style sea-level oscillation of 30-50 m with similar maxima for each subsequent rise. These oscillations are interpreted as glacio-eustatic in origin, related to fluctuations in southern Gondwanaland continental ice sheets, but of a much smaller size than ice sheets formed in the Pleistocene. These late Mississippian successions provide an example of multiple palaeokarsts separated by predominantly subtidal deposits, reflecting ice-enhanced, eccentricity-generated eustatic cycles.

In summary palaeokarst development can be related to major sequence (and seismic) stratigraphic boundaries but it can also be related to higher frequency sea-level changes. In the latter case the role of major continental ice build-ups is a crucial factor in controlling, not only the degree of karstification, but also the architecture and regional diagenesis of the carbonate platform.

However, palaeokarsts formed in association with the higher frequency sea-level falls are, as one might expect, generally less well developed than those where 'continental' processes operated (Chapter 1). The degree of karstification is related to the length of exposure, climate, relief and the balance between uplift and erosion. Major palaeokarst development is usually associated with some tectonic control. For example Palmer and Palmer (1990) note that the two major palaeokarsts in the USA are related to unconformities reflecting major tectonics. These cap the Sauk sequence (Early/Middle Ordovician) and the Kaskaskia sequence (mid-Carboniferous). This association of palaeokarst and major tectonic phases is discussed in Chapter 3. Even the study of more local palaeokarsts consistently shows that major karstic porosity development is usually related to a tectonic control. Some of the following sections illustrate these karst-tectonic associations.

2.6 PALAEOKARST ASSOCIATIONS

Rather than review palaeokarst case studies, it might be instructive to chose some distinctive associations with direct relevance to general exploration in carbonate terrains. To this end, palaeokarst associated with three typical play associations will be discussed: reefs, evaporite basins and platform margins. In addition, aspects of palaeokarst development in foreland and extensional basins will also be reviewed.

2.6.1 Reefs and Karst

The importance of karstic processes and reefs has been noted by Purdy (1974 a,b), who argued that such processes are important, during low stands, in generating relief used for later reef growth during rising sea-levels. The karst relief acted as antecedent topography, where preferential reef growth occurs. As a result, reef growth can be localised on original karst features, especially around the margins of isolated platforms or shelves to create barrier reefs or atolls.

Isolated reef atolls, and also any isolated carbonate build-up, such as a small platform, have the potential to develop distinctive groundwater geometries. On such isolated settings, following subaerial exposure, the meteoric lens will float on the denser, marine phreatic beneath and under the island. In unconfined aquifers, the meteoric phreatic lens extends below sea-level approximately forty times the height of the water-table above sea level, known as the Ghyben-Herzberg relationship (Todd, 1980). This is an important effect by which a deep meteoric phreatic lens can develop following relatively small sea level falls.
This simple relationship assumes no mixing of the marine and meteoric waters which does take place. The relationship does apply to lenses bounded by the 50% isochlor surface (Vacher, 1978), but in strongly heterogeneous aquifers such as reefs, the lens has a more complex geometry (Buddemeier and Oberdorfer, 1986). The small catchment areas and low relief on such islands will not result in extensive karst development, but meteoric diagenesis can be significant.

A more unusual relationship between karst and reefs may exist for some types of karst features resemble pinnacle reefs. Mogotes (tower karst) are large, isolated pinnacles of limestone often with considerable relief, and some striking examples have been documented from Palaeozoic palaeokarsts. Bless et al. (1980) and Poty (1980) have described Devonian tower karst from Belgium, which was progressively drowned during the early Carboniferous. Locally, the mogotes even acted as a substrate for coral growth. Mogotes associated with extensive mineralization have also been described from Colorado by Maslyn (1977) and De Voto (1988). In this example, Devonian and Mississippian shallow-water limestones were heavily karsted to form tower karst with 30 m of relief, and also as ridges 120 m long. These were buried by Pennsylvanian organic-rich shales. Similar karst has been recorded from Ordovician sequences of Quebec by Desrochers & James (1988).

In all these cases of ancient mogotes, one is left wondering if, given greater relief, such structures might not be interpreted, on seismic profiles, as reefs and not palaeokarst. It is reasonable to ask if some reefs, recognised from purely seismic data, may not actually be mogote karst, perhaps developed on non-reefal limestone?

### 2.6.2 Palaeokarst and Evaporites

While the occurrence of dissolved evaporites and interstratal karst has been recorded on many occasions, there is another association between karst and evaporite basins which requires consideration. During the desiccation of large salt basins the down-draw of saline brines causes a regional lowering of the groundwaters. For example, during the Messinian desiccation of the Mediterranean the fall in the base-level was over 2000 m and lasted for as long as one million years. As a result, major valleys and gorges were cut and deep cave systems formed (Corra, 1986). Recently Kendall and Harwood (1989) have suggested that some of the Castile evaporites of the Delaware Basin formed in very shallow brine pools and not in a deep saline basin as had been previously thought. Such an interpretation requires a drop in sea-level in the basin of probably over 500 m, and raises the possibility that deep karstification could have taken place in the surrounding platform sequences (Fig. 2.16). The extent of karstification will have depended on the climate (which must have been one of low rainfall) and length of exposure. It is possible that deep palaeokarst systems may develop around salt basins during their desiccation, but as yet little work has been carried out on such systems. Since many major hydrocarbon provinces are associated with saline basins (Devonian of Alberta, Williston Basin, Zechstein of Europe, Infra-cambrian of Oman), the possibility of major karstic porosity formation around the basin margins should be considered.

### 2.6.3 Palaeokarst at Platform/Shelf Margins

Whereas, broadly speaking, palaeokarst evolution under a given set of circumstances develops along similar pathways in its earliest stages, regardless of platform configuration (ramp, shelf), there is a particular style of karstification associated with steep (aggrading or stationary) escarpment margins that is worthy of special consideration for such fossil margins are traditionally targets for exploration because of their association with grainstones or reefs.

The steep margins of some platforms and shelves, typically a function of high rates of sea-level rise or structural control, are modified by large scale collapse. Basically the margins spall-off, because of lateral unloading,
Fig. 2.16 Effects of desiccation on karst development in a salt basin, e.g. Delaware Basin (modified from Kendall and Harwood 1989).
Fig. 2.17 Schematic representation of a typical platform margin fracture system. During low stands karstification will take place. During high stands marine sediments are deposited in the fractures. The fractures may connect both interior mixing zone karst systems and other porosity zones within the platform.
(Freeman-Lynde et al., 1981; Mullins and Neumann, 1979), and this manifests itself on the margins as a series of fractures. Along the margins of the present-day Great Bahama Bank, these fractures are often multiple and complex, sometimes curvilinear and near vertical, but showing no displacement (Smart et al., 1988a). Many can be traced laterally for tens of kilometres, both on the submarine banks and across land (e.g. Andros Island) (Dougherty et al., 1986). Major perpendicular joint sets are also present. On land, both sets of fractures are enlarged by dissolution processes giving straight but irregular rifts up to half a metre wide and up to several metres deep. Some fractures, paralleling the bank margin, have been greatly enlarged and constitute major cavities, averaging 2-5 m wide (but up to 30 m), in excess of 90 m deep and running for hundreds of metres (Smart et al., 1988a). The development of the fractures, in the case of the Bahamian platforms today, appears to be due to lateral unloading, in other examples the margins may be directly tectonically defined and the fractures reflect the regional tectonics (Vera et al., 1988), although again lateral unloading is important. In other cases, differential compaction of the basin and platform has created stresses along the margins (Playford, 1984).

Similar fracture systems have been widely documented from ancient platform sequences. Many of these ancient fractures have been filled with marine sediment to give "neptunean dykes". Some of these fractures also show evidence of karstification. Platform fracture systems have been described from the Devonian Canning Basin (Playford, 1984), and the Tarfita platforms of Morocco (Wendte et al., 1984). They are especially well documented from the Tethyan break-up sequences of southern Europe, and Vera et al. (1988) provide a detailed description of Jurassic platform margin palaeokarsts in southern Spain (Fig. 2.18). Smart et al. (1988a) have reviewed the terminology for such fracture systems and their deposits.

The exploration significance for this type of palaeokarst is clear. At high-relief platform margins extensive fracture systems, with or without palaeokarst, should be very common. Many platform margins in the Tethyan sequences are defined by listric or normal faults and footwall uplift is a common process at such zones, resulting in subaerial exposure and karstification (discussed in next section).

There are other implications for this model. The fracture systems in the Bahamas connect with interior platform cavity systems related to mixing zone dissolution (Smart et al., 1988b). Indeed, a situation could be envisaged where platform margin fracture systems may connect with earlier "stacked karst" within the platform interior providing a large interconnected reservoir (Fig. 2.17).

The model offered here (Fig. 2.17) recognises two major controls on karst development: the platform parallel fracture system, and the development of mixing zones. The karst hydrological system is effectively an autogenic one. However, in the case of shelves bordering land masses, or carbonate ramps, allogetic hydrologic systems are likely to arise with more complex karst development. In such cases, very extensive karstification may result including mixing zone dissolution evolution (Hanshaw and Back, 1980; Back et al., 1984).

2.6.4 Paleokarsts in Extensional Tectonic Regimes

Recent studies of Triassic extensional tectonic systems in Europe have revealed distinctive palaeokarst developments. These studies may serve as a useful guide to exploration in similar tectonic settings.

The external Subbetics of southern Spain contain a sequence of classic Tethyan break-up sequences in the Granada-Jaen area (Molina et al., 1985). Extensive shallow water, lower and Middle Jurassic platforms were dissected by listric faults and subsequently rapidly subsided so that platform sequences are capped by pelagic limestones. Initially subaerial exposure of the platforms occurred killing off carbonate production and extensive fracture systems formed along the faulted tilt-block margins, as described in the previous section. These
Fig. 2.18 Schematic reconstruction of the evolution of the carbonate platform of the External Subbetic region of southern Spain. Note the preferential karstification on the crests of the hanging walls. Based on Molina et al. 1985.
were locally karstified and were later filled by speleothems and various pelagic sediments (Vera et al., 1988). By detailed mapping of the region, it has been possible to recognize differences in the degrees of karstification across the various tilted blocks (Fig. 2.18). The footwall zones of each block were preferentially karsted, with the degree of dissolution decreasing down the adjacent hanging wall (Fig. 2.18). This is interpreted as reflecting footwall uplift. Some blocks were never subaerially exposed, while others were exposed for short periods of time and relatively lightly karstified. Some blocks remained exposed until the Upper Cretaceous. In these latter examples, extensive cavern development and collapse took place (Vera et al., 1984). More, as yet unpublished examples of this style of karstification also occur in the Mesozoic Lusitanian Basin of Portugal, associated with Atlantic opening.

As a general rule, such palaeokarst systems should be relatively small, capping individual tilt-blocks. Nearby blocks may lack such karsts. Such a setting would result in buried-hill-type traps and should be readily recognizable from seismic data. The adjacent half-graben basins may well have been the sites of organic-rich sediment accumulation with anoxia, as was the case with many of the Tethyan and the Lower-Middle Jurassic systems.

In the examples noted above, the karst developed as rift-onset unconformities early in passive margin growth. However, more extensive exposure occurs later in passive margin development with regional 'break-up' unconformities.

2.6.6 Palaeokarst and Foreland Basins

Palaeokarsts are also a typical feature of many foreland basin sequences. They occur in carbonate sequences, related to the formation of peripheral bulges (Fig. 2.19). As a consequence of flexural downwarp due to thrust emplacement and loading, uplift occurs in the foreland with subaerial exposure (Quinlan and Beaumont, 1984). Such peripheral bulge palaeokarsts have been recorded from the Ordovician of the Canadian Appalachians by Desrocher and James (1988) and James et al. (1989). Similar associations have also been noted in the Tethyan sequences of the northern Mediterranean by Bosellini (1989). One example has already been noted, that of the Carboniferous Limestones of South Wales (Figs. 2.7 and 2.8). In this example, the northern limb of the synclinorium represents the peripheral bulge region, where uplift and karstification occurred in pre-Namurian (pre-Pennsylvanian) times. To the south, rapid subsidence occurred taking the limestones down to 7 km locally. The most northern areas have not matured, while thick shales developed and matured in the foredeep area.

The significance of such settings is that palaeokarst might be regarded as an integral part of foreland basin development. With propagation of the bulge ahead of the migrating thrust front extensive karsted areas may form.

2.7 SUMMARY

Most limestones are deposited in very shallow water and are susceptible to exposure during sea-level changes. Both meteoric and mixing zone dissolution diagenesis should be common features in such limestones. During burial diagenesis and partial exhumation, limestones may be affected by dissolution related to aggressive fluids derived from deeper settings.

Whereas there have been relatively few studies of palaeokarst systems, some broad patterns are emerging, and distinctive palaeokarst facies and palaeokarst associations can be recognized, which should provide useful guides in hydrocarbon exploration.

The present-day situation is not necessarily a reasonable guide to karst development in the past. Changes in limestone composition and in global climate may have reduced or enhanced the potential for karst formation at various times in the past.

It is already clear that many major palaeokarsts are polyphase in origin. The possibility of similar situations must be borne in mind when attempts are made to interpret any palaeokarst system.
Recognizing and interpreting palaeokarsts in the subsurface is difficult but a variety of techniques can be applied. Modelling a palaeokarstic reservoir is particularly difficult and it is likely that some fractured reservoirs may be palaeokarstic in origin.

Peripheral Bulge Region

Fig. 2.19 Schematic diagram showing the formation of palaeokarst development in foreland basins.
CHAPTER 3
PALAEOKARST: PRACTICAL APPLICATIONS
by M. Esteban

3.1 INTRODUCTION

It is commonly accepted that at least 20-30% of recoverable hydrocarbons appear in some way to be related to unconformities (e.g. Weeks, 1952; Gibson 1965; Levensen, 1954; Moody et al. 1970). This percentage is probably much higher for carbonate formations affected by unconformities. The processes that form unconformities or affect their character can enhance, modify or obliterate carbonate porosity and permeability. In addition to the development of reservoirs, unconformities can also be directly involved in migration pathways and trapping. Although the importance of unconformities in exploration and production is well established, it is important to realize that our knowledge about them is very limited when compared to faults, reefs, deltas or many other geologic features of economic significance.

Some statistical accounts of international oil and gas fields fail to recognize any significant association with unconformities (Halbouty, 1970). Several factors may contribute to this perception. First of all, most classifications only consider as unconformity-related those reservoirs and traps adjacent to the unconformity. However, as later discussed in this chapter, some reservoirs and traps located many hundreds of feet below the unconformity were generated by the processes that formed the unconformity (Fig. 3.1). In other cases, reservoirs are classified as "fracture type" without assessing the possibility that the reservoir properties were enhanced by the same fluids involved in the formation of the unconformity. Furthermore, in other instances the unconformity itself may be very subtle or difficult to recognize with the standard techniques. Early exploration delineated some positive carbonate features, and these were described as "reefs" although an organic buildup was not clearly recognized. Some of these (Golden Lane, Scurry County and some pinnacle reefs, (Fig. 3.2) are now considered or debated as erosional features underlying unconformities, associated with or independant of carbonate shoals or reefs. Finally, many traps classified as palaeogeomorphic are in fact features produced by the same processes that formed the unconformities. In summary, when the unconformity-reservoir-trap relationship is viewed from a genetic perspective (rather than purely in terms of position or spatial proximity) it becomes a key factor in understanding the occurrence of many large hydrocarbon accumulations.

Carbonate reservoirs and traps may show different relationships to unconformities. In many carbonate reservoirs the bulk of the reservoir porosity may have been developed or destroyed by unconformity-forming processes. In other reservoirs most of the porosity was generated by other means (primary deposition, tectonic fracturing, deep burial vuggy corrosion) and only modified to some degree by the unconformity-forming processes. The assessment of the role of unconformities in carbonate reservoirs requires an evaluation of the inherited porosity and the amount of porosity generated by those processes.

3.2 KARST AND REEFS: PRESENT SITUATION

Palaeokarst reservoirs, as a particular element in unconformity plays in carbonates, commonly have presented a somewhat confusing picture in exploration and production. Many of these reservoirs appear as rather unpredictable, chaotic and heterogeneous. There is a vast amount of karst literature, but most of it is not directly applicable to exploration and production problems because of the emphasis on geomorphology, hydrology or sportive speleology. The present situation is very similar to reef exploration and production in the early
Fig. 3.1 Schematic cross-section of karstic terrain. Note that potential karst targets are not restricted to a zone adjacent to the unconformity.
Fig. 3.2 Isopach map of Horseshoe Reef Complex, West Texas (A) (Vest, 1970). Seismic cross-section of faulted platform rim, with relief produced by karst processes (B). Both illustrate the types of features that have generally been interpreted as reefal but are now considered to be strongly modified reefal features or features produced by erosion associated with unconformities.
1950's. Available literature at the time (mostly from marine biology and descriptive palaeontology) was not suitable for the formulation of adequate exploration and production models. It was necessary for the industry to examine modern and fossil reefs extensively, in order to provide applicable and successful models. Unconformities, palaeokarst and reefs can all be detected by seismic methods (Fig. 3.2), but they also can be very subtle, misinterpreted or ignored. Their role in the development of reservoirs can be exaggerated, underrated or confused; there can be problems of semantics, recognition of different types and prediction of reservoir trends. As with reefs, palaeokarst plays can be as diverse as the different types of unconformities and the carbonate structures and lithologies in which they are developed.

The understanding of carbonate porosity and diagenesis in exploration and production has been dominated by the application of models developed in the study of the Pleistocene and recent carbonates, mostly in the Caribbean area. However, this Caribbean model offers many limitations when applied to a large variety of economically important unconformities; namely, those involving large time gaps (Table 3.1). Choquette and James (1988) have addressed these different karst types.

3.3 TYPES OF UNCONFORMITIES

An unconformity implies a stratigraphic discontinuity or gap (lacune) which represents chronostratigraphic intervals missing through nondeposition (hiatus) and/or lithostratigraphic intervals missing through erosional truncation. An unconformity implies that significant (measurable) time and/or stratigraphic section is missing at the unconformity surface. "Measurable" is the key concept, because it depends on our limits of resolution and powers of observation. There are minor stratigraphic discontinuities (diastems) without demonstrable or measurable stratigraphic gap. Many bedding planes in conformable (concordant) sedimentary sections are diastems showing evidence of minor depositional discontinuities produced by erosion, weathering, and nondeposition. In a regional study of an unconformity (Fig. 3.3), typically major stratigraphic gaps and angularities observed along basin margins and on local high points toward the basin center into paraconformities and diastems (conformity and concordance).

The chronostratigraphic representation of an unconformity on a structure (Fig. 3.4) shows:

1) Minimum gap: the time interval not represented by the sedimentary record in the area, either by complete erosional removal or by nondeposition. The minimum gap corresponds to the difference between the youngest age of the truncated section and the oldest age of the onlapping section.

2) Maximum gap: the maximum time interval absent in the sedimentary record in the area. The maximum gap corresponds to the difference between the age of the truncated section and the age of the youngest bed of the onlapping section.

This method of chronostratigraphic representation of an unconformity allows interregional comparison of unconformities. This will be applied to the palaeokarst case histories in chapter 4.

In relation to the magnitude of the stratigraphic gap, a hierarchic classification of stratigraphic discontinuities is proposed here (Fig. 3.5). The local stratigraphic discontinuities without measurable stratigraphic gaps (conformities) correspond to bedding planes (diastems) or to parasequence boundaries (discontinuities capping shoaling-upward cycles). Unconformities can be classified as single or composite. Single unconformities represent stratigraphic gaps of approximately one biozone and the sequence boundaries (3rd order cycles) of Vail et al. (1977). Composite unconformities are formed by the stacking or superposition of single unconformities, with time gaps corresponding to stages (super-
Fig. 3.3 Traditional classification of unconformities and their common regional pattern.

Table 3.1 Subaerial diagenesis: Caribbean model versus a general model.
Fig. 3.4  Chronostratigraphic representation of an unconformity.
Fig. 3.5  *Hierarchy of stratigraphic discontinuities.*
unconformities), series, systems (mega-unconformity) or erathems. Figure 3.6, which shows the hierarchy of stratigraphic discontinuities with their approximate time gap in millions of years and corresponding stratigraphic units, attempts to integrate widely used terms in sequence stratigraphy. It differs substantially from the proposed palaeokarst classification of Choquette and James (1988).

A large number of palaeokarst case histories are associated with composite unconformities (Fig. 3.6). Cases of palaeokarst production related to single unconformities (sequence boundaries, 3rd order cycles) are significantly less frequent, although there are important examples (e.g. Yates Field, Craig, 1988). Palaeokarst associated with parasequence boundaries (discontinuities capping shoaling-upward cycles) is also very common. Statistical comparison of reserves to type of stratigraphic discontinuity is considered to be inconclusive because of the importance of other factors, such as size of structure, seal, source-rock charge etc.. Palaeokarst of the Caribbean-type (unstable carbonate mineralogy, primary porosity, interparticle flow, etc.) is clearly applicable to the cases related to parasequence boundaries (depositional palaeokarst of Choquette and James, 1988). Palaeokarst reservoirs related to composite unconformities (interregional palaeokarst of Choquette and James, 1988) follow non-Caribbean models (stabilized mineralogy, predominance of secondary porosity, minor or no interparticle flow, etc.) and they will be emphasized in this chapter. Palaeokarst at single unconformities (3rd order sequence boundaries) may present a combination of Caribbean and non-Caribbean karst types and may explain the high production rates of some of them. Syntectonic unconformities (progressive discordances) show no major palaeokarst association. This is probably due to the predominance of siliciclastic sedimentation in these settings.

Subaerial exposure can affect carbonate sediments and rocks immediately or very shortly after their formation (Caribbean-type diagenesis, syngenic or eogenetic weathering). However, in many circumstances, subaerial exposure processes affect much older carbonate rocks and sediments that have suffered intermediate or deep burial, and may have been deformed by tectonic events. This implies at least one cycle of regional subsidence and uplift with dismantling of the sedimentary cover and requires a long period of exposure. Immediately upon subaerial exposure carbonate formations are affected by meteoric diagenesis. Some of these processes involving surface and subsurface meteoric waters produce recognizable surface and subsurface landscapes known as karst by geographers and hydrologists. Karst is defined as the product of subaerial exposure in carbonates consisting of an integrated drainage system including conduit flow. This drainage system is formed by dissolution and mechanical erosion of carbonates by meteoric waters (and their mixing with marine, hydrothermal, and compaction waters) enhancing pre-existing permeability networks (fractures, bedding planes, primary porosity and secondary porosity). Karst can occur in: (i) terrestrial exposure environments away from coastal influences, (ii) coastal exposure environments (sea or lake), (iii) beneath the sea, many meters below the zone of coastal exposure (submarine karst), and (iv), hydrothermal environments.

Similarly, as with reefs in the 1950's, explorationists are confronted with a multitude of definitions, terms and concepts in relation to karst. In this chapter, two main types of controls influencing karst reservoirs are considered (Fig.3.7). Some karst reservoirs were formed by an early stage of subaerial exposure, and the resulting reservoir properties tend to show a marked control by depositional facies patterns and lithologies (Caribbean model). This type of karst reservoir grades into the depositional carbonate reservoirs, which are dominated by primary porosity or by early subaerial exposure without the development of an integrated drainage system with conduit flow. Other karst reservoirs were formed by late subaerial exposure (after a phase of burial) and commonly have a strong
<table>
<thead>
<tr>
<th>STRATIGRAPHIC DISCONTINUITIES</th>
<th>ORDER</th>
<th>TIME GAP SCALE</th>
<th>CORRESPONDING STRAT. UNIT</th>
<th>ASSOCIATED RESERVOIR</th>
<th>FIELD EXAMPLES</th>
</tr>
</thead>
<tbody>
<tr>
<td>E. UNCONFORMITY</td>
<td>1st</td>
<td>200 My</td>
<td>ERATHEM</td>
<td>MEGASEQUENCE 200 My</td>
<td>RENGIU</td>
</tr>
<tr>
<td>MEGA-UNCONFORMITY</td>
<td></td>
<td>&gt;60 My</td>
<td>SYSTEM</td>
<td>SUPERSEQUENCE SET 400 My</td>
<td>ARGYLL</td>
</tr>
<tr>
<td>S.U. SET</td>
<td></td>
<td>30 My</td>
<td>SERIES</td>
<td>SUPERSEQUENCE (sequence of Sloss) 10-100 My</td>
<td>GOLDEN LANE</td>
</tr>
<tr>
<td>SUPER-UNCONFORMITY</td>
<td>2nd</td>
<td>4-12 My</td>
<td>STAGE</td>
<td>BIOZONE</td>
<td>CASABLANCA</td>
</tr>
<tr>
<td>UNCONFORMITIES</td>
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<td>~1 My</td>
<td>DEPOSITIONAL mesothem</td>
<td>SEQUENCE 1-10 Type 1</td>
<td>VEGA</td>
</tr>
<tr>
<td>COMPOSITE</td>
<td></td>
<td>~1 My</td>
<td></td>
<td></td>
<td>DEVONIAN REEFS</td>
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<td>3rd</td>
<td>0.01-1 Variable</td>
<td>ROTATIONAL syntectonic</td>
<td>PARASEQUENCE cyclothem 0.2-0.5</td>
<td>YATES</td>
</tr>
<tr>
<td>(sequence boundaries)</td>
<td></td>
<td></td>
<td>ONLAP-OFFLAP cum. wedge</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TYPE 1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>TYPE 2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SYNTECTONIC UNCONFORMITY</td>
<td>3-4</td>
<td>0.01-1 Variable</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>BOUNDARY OF SHOALING CYCLES</td>
<td>4th</td>
<td>0.01</td>
<td></td>
<td></td>
<td>SMACKOVER</td>
</tr>
<tr>
<td>BEDDING PLANE</td>
<td>5th</td>
<td>0.001</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Fig. 3.6 Stratigraphic discontinuities, time gaps and types of reservoir.
Spectrum of karst reservoirs, their controls and relationships with other types of carbonate reservoirs.

<table>
<thead>
<tr>
<th>DEPOSITION</th>
<th>EARLY SUBAERIAL EXPOSURE</th>
<th>BURIAL</th>
<th>LATE SUBAERIAL EXPOSURE</th>
<th>BURIAL</th>
</tr>
</thead>
<tbody>
<tr>
<td>Compaction</td>
<td>Early fractures, collapse</td>
<td>Pre-unconformity fractures, unloading, tectonic events</td>
<td>Post-unconformity fractures and tectonic events</td>
<td></td>
</tr>
</tbody>
</table>

Depositional control

Structural control

HYDROTHERMAL KARST

KARST RESERVOIRS

<table>
<thead>
<tr>
<th>DEPOSITIONAL RESERVOIRS</th>
<th>FRACTURED RESERVOIRS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Motion (New Mexico)</td>
<td>Fractured (Texas)</td>
</tr>
<tr>
<td>Fateh Qai (Arabian Gulf)</td>
<td>Fractured (Texas)</td>
</tr>
<tr>
<td>Yates (Texas)</td>
<td>Fractured (Texas)</td>
</tr>
<tr>
<td>Euk-A TREE (U.K)</td>
<td>Fractured (Texas)</td>
</tr>
<tr>
<td>Golden Lane (Mexico)</td>
<td>Fractured (Texas)</td>
</tr>
<tr>
<td>Casbiania (Spain)</td>
<td>Fractured (Texas)</td>
</tr>
<tr>
<td>La Paz (Venezuela)</td>
<td>Fractured (Texas)</td>
</tr>
<tr>
<td>Paris Basin (France)</td>
<td>Fractured (Texas)</td>
</tr>
<tr>
<td>Smackover (Texas)</td>
<td>Fractured (Texas)</td>
</tr>
<tr>
<td>Scurry County (Texas)</td>
<td>Fractured (Texas)</td>
</tr>
<tr>
<td>Elk Point (Wyoming)</td>
<td>Fractured (Texas)</td>
</tr>
<tr>
<td>Ul (Texas)</td>
<td>Fractured (Texas)</td>
</tr>
<tr>
<td>Fathab (Canada)</td>
<td>Fractured (Texas)</td>
</tr>
</tbody>
</table>

FIELD EXAMPLES

Monterey Fm (California)

La Paz (Venezuela)

Casbiania (Spain)

Belle River (Canada)

Scurry County (Texas)

Golden Lane (Mexico)

Elk Point (Wyoming)

Fateh Qai (Arabian Gulf)

Motion (New Mexico)
structural control (fractures, bedding planes) of karst porosity. This type of karst reservoir grades into fractured reservoirs when late subaerial exposure processes do not result in a fully integrated drainage system with conduit flow. Within this perspective there are considered borderline or transitional cases at both ends of the spectrum of karst reservoirs. Hydrothermal karst is a special case of karst reservoir that can occur as a late overprint on pre-existing palaeokarst systems or alternatively can generate the bulk of the karst porosity. Hydrothermal karst reservoirs can occur during burial and are controlled by post-unconformity features; hydrothermal karst reservoirs can also grade into reservoirs formed in burial environments.

3.4 RESERVOIR DEVELOPMENT AND TRAP CONFIGURATION

Subaerial exposure processes that generate unconformities (and palaeokarst) in carbonate formations may develop different geometric styles of reservoirs and traps. These are summarized in Enclosure 3.1, which shows the five basic styles and nine variations (Esteban-Wilson Classification). These schematic diagrams show no horizontal or vertical scales. The dotted area represented as the reservoir in Enclosure 3.1 does not imply homogeneous reservoir properties; they may include significant non-reservoir rock volumes. These unconformity reservoirs and associated palaeokarst show the following general features:

1) Blankets over broad arches. The associated unconformities show gentle palaeotopography over wide regional arches with low structural complexity, typically corresponding to Type 2 or Type 1 sequence boundaries of Vail et al. (1984) and involving regional dolomitization, microporosity and leaching; karst zones and conduits are commonly poorly developed. One example is the top of the Siluro-Devonian Hunton Group under the pre-Frasnian Woodford unconformity over the Midcontinent, USA. Other examples of this style are the top of the Aptian Shuaiba carbonates in the Hasa coast of Saudi Arabia and the top of the Edwards (Lower Cretaceous) over the San Marcos platform, Texas, with extensive dolomitization and leaching during Albian time.

2) Mosaics of remnant highs. These are non-linear patterns that correspond to two general sub-styles:

(A) Depositional-erosional. Highs under Type 1 and some Type 2 sequence boundaries, including undifferentiated erosional features (Lost Soldier, Elk Point, Montana; Campeche-Reforma in Mexico) or patch-reefs, knobs of cemented grainstone shoals, mounds of different types (Nena Lucia, Scurry County, Texas), with significant modification by the unconformity. This pattern can also be related to diapiric domes or to intensively and repeatedly deformed areas with crossing trends of faults or folds (Casablanca-Amposta, Spain). This type of reservoir commonly contains breccia porosity and well developed integrated karst conduits. This is a very common style of reservoir that locally grades into linear trends.

(B) Stacked convergent exposure surfaces on slightly emergent mounds, shoals or patch-reefs. These surfaces correspond to boundaries of parasequences (shoaling-upward cycles). Large cavern porosity is uncommon. Exposure times are relatively short and the diagenetic evolution follows the Caribbean model. Some of the Arabian Gulf fields (Fahud) and the pinnacle reefs of the Devonian of Canada correspond to this type.

3) Linear-trends. Predominant linear structure associated with Type 1 and 2 sequence boundaries. Breccia and cavern porosity is very common, with well developed and deep integrated karst. This style can be associated with hydrothermal karst, in which reservoirs are preferentially developed in lows. There are three major sub-styles:

(A) Fault-controlled. Linear structure may develop before, during or after the exposure event. This includes the potential for the deepest reservoirs of all
the styles. Hydrothermal modifications are common e.g. Albion-Scipio (Michigan), Renqiu (China).

(B) Anticline-controlled. Linear structure developed before or during exposure event. There is good potential for artesian karst systems e.g. Kirkuk (Iraq).

(C) Depositional-erosional. These occur along platform-margin rims that can develop as rim-effect erosional features or controlled by marginal reefs, shoals or mounds significantly modified by the unconformity. They can also be associated with fault-controlled trends e.g. Golden Lane (Mexico), Vega (Sicily), Puckette (Texas).

4) Wedges and mantles above unconformity. These are sediments related to residual deposits or to an onlapping sequence produced either by the unconformity forming processes or by successive deposition. Typically these wedges are associated with flanks of Styles 2 and 3 features and to late stages of karst evolution. Breccias are very common, and this style of reservoir may contain a wide variety of lithologies. This includes debris-flow breccia, chert residues and regoliths, coastal cliff deposits, subaerial scree fans, etc.) with variable amounts of siliciclastic deposits. Reservoirs may be formed entirely by submarine exposure processes in some settings e.g. Poza Rica (Mexico).

5) Layers of shallowing cycles. Multiple shallowing-upward hemicycles (parasequences) with intermittent short exposure during or shortly after deposition; Caribbean-type of diagenetic evolution with poorly developed cavern porosity and typically with multiple and thin reservoir horizons. There are two main sub-styles:

(A) Layer-cake. e.g. Jurassic-Lower Cretaceous in the Arabian Gulf (Qatif, Fateh); Jurassic in Paris basin; Fairway, Texas.

(B) Layer-mound. With low growth rates, these carbonate mounds or buildups may grade into Style 2B. Michigan basin pinnacle reefs; Pennsylvanian algal mounds (USA); Miocene reefs (Far East).

This classification of reservoir styles is independent of the carbonate lithologies involved. Carbonates developed as Styles 1, 3 and 5 may include evaporite intercalations. Some fields may contain different styles, and the scale of the structures in observation may influence the classification. For instance, in the Gulf of Valencia the large-scale reservoir styles are dominated by linear rift trends, but the local fields or parts of fields are mosaics of remnant highs produced by the intersection of complex patterns of rift block faults.

3.5 ASSESSMENT OF PALAEOKARST PLAYS

3.5.1 Palaeokarst occurrence

Karst does not occur randomly through time and space; there are geologic time intervals in which karst occurrence is more widespread and certain geologic settings which are more conducive to different types of karst reservoirs. As a consequence, the assessment of palaeokarst plays requires a preliminary screening of those regional scale conditions favorable to karst reservoir development. This preliminary screening involves: i) delimitation of carbonate provinces, ii) consideration of palaeogeographic and palaeoclimatic patterns and development of different types of karst, iii) consideration of relative oscillations of sea-level on potential karst profiles.

Following preliminary screening, the assessment of potential palaeokarst requires analysis of local geology at the scale of drillable structures, with consideration of: i) lithologies and structure of the unconformity subcrop, ii) geologic history of the unconformity, with evolution of tectonic events, palaeogeomorphology, paleohydrology, karst facies, and iii) transgressive stratigraphy and burial of the paleokarst. This unconformity-palaeokarst analysis requires input of local geological and geophysical data.
Fig. 3.8. World distribution of modern karst (Balass, 1962).
Modern karst is present in a wide range of environmental settings from subarctic to tropical (Fig. 3.8); however, many descriptions of palaeokarst in the literature seem to be restricted to warm temperate, to semiarid and to subtropical environments. This is particularly true when considering palaeokarsts with hydrocarbon production. Important exceptions to this are palaeokarsts with bauxites, red beds and laterites in high altitudes that preclude any tropical influence (as discussed below). Reasons for the environmental restriction are not known, although it may, in part, be due to misinterpretation of the exposure environments.

The palaeoclimatic and palaeogeographic reconstructions of the Chicago Group (Ziegler et al, 1981a and 1981b) and Habicht (1979) are the basis for the reconstructions of Figures 3.9 - 12. Key climatic indicators are included: distribution of carbonate provinces, evaporites, bauxites, red beds and coal deposits. Known and possible palaeokarst settings are added for each time interval. Where palaeokarst occurs at composite unconformities with large time gaps (superunconformities, megaunconformities), the last karst event is considered the most relevant in terms of reservoir properties and is plotted on these maps.

A number of general trends emerge from these maps. Many palaeokarsts occur within the palaeotropics, in association with red beds and evaporites and away from the main accumulations of coal deposits. Coal deposition implies high organic productivity and low rates of degradation, which is often associated with areas with poor drainage. This poor drainage may explain the lack of good karst development. After the Carboniferous, areas with predominant coal deposition shifted toward higher latitudes.

It is also very significant that palaeokarst occurred in coastal areas rather than in continental interiors. During Pangea-type of continental assemblages (Late Carboniferous to Triassic), movement of palaeokarst occurrences was toward higher latitudes, with peripheral palaeokarst development (even associated with coal deposits in some of areas).

The Miocene reconstruction corresponds to the neokarst realm and the origin of many modern karst systems. The Miocene is considered as the beginning of a shift of karst development (neokarst) toward higher latitudes, culminating in present-day karst distribution (Fig. 3.8). Quantitative models of rainfall distribution for the Mesozoic and the Tertiary (Parrish et al, 1982a and 1982b) are reproduced with the location of palaeokarst occurrences.

Another general trend which emerges from these figures shows that in the Early Triassic palaeokarst was associated with areas of predicted high rainfall. Following the breakup of Pangea, the distribution of karst was effectively enlarged to include areas of moderately high rainfall (Late Jurassic) and moderately low rainfall (Cenomanian-Miocene).

While many palaeokarsts are interpreted to have developed within the tropics, there are a significant number of departures from the general trend. Some of these exceptions are due to different palaeogeographic reconstructions; others may imply climatic patterns which are significantly different from reconstructed models. Some palaeogeographic models may be biased by erroneous interpretations of palaeoclimatic indicators. For instance, thick carbonate platforms are not restricted to tropical settings. Reefal (ahermatypic), skeletal-rich carbonates are also very abundant in temperate waters.

In summary, refined palaeogeographical and petrological analysis may provide evidence for a variety of non-tropical karst occurrences. However, based on current data, it appears that hydrocarbon-producing palaeokarst is distributed within palaeo-tropical zones.

3.5.2 Palaeokarst and relative sea-level oscillations

The curves of global relative changes of coastal onlap (Haq et al, 1987; Ross and Ross, 1987; Johnson et al, 1985, and other sources) for the Palaeozoic (Fig 3.13) and the Mesozoic-Cenozoic provide a framework for distinguishing two main
Fig. 3.9  Ordovician palaeogeography and palaeokarst.  A. Mid Ordovician reconstruction.  K = known palaeokarst.  Climatic zonation indicated by symbols: circle = tropical; square = temperate; diamond = polar.  Ocean currents indicated by arrows: thin = cold; thick = warm.  Shading indicates relief: dark = mountains to white = basins (Ziegler et al., 1981).  B. Late Ordovician reconstruction.  Sediments of different Ordovician ages are plotted with Late Ordovician palaeogeography (Scotese et al., 1979).
Fig. 3.10 Carboniferous palaeogeography and palaeokarst. A. Late-Early Carboniferous reconstruction. Climatic zonation indicated by symbols: circle = tropical; triangle = subtropical; square = temperate. Ocean currents indicated by arrows: thin = cold; thick = warm. Shading indicates relief: dark = mountains to white = basins (Ziegler et al. 1981). B. Late Carboniferous reconstruction. Sediments of different Carboniferous ages are plotted with Late Carboniferous palaeogeography (Scotese et al. 1979).
Fig. 3.11 Triassic palaeogeography and palaeokarst. A. Early Triassic (Induan) reconstruction. K = known palaeokarst. Contours of predicted rainfall distribution. Numbers show relative values only (no units are implied). Low rainfall = <50; moderate rainfall = 100-200; high rainfall = >200. Shading indicates relief: light = continental shelf; medium = lowlands; dark = highlands (Parrish et al., 1982). B. Sediments of different Triassic ages are plotted with Early Triassic palaeogeography (Scotese et al., 1979).
Fig. 3.12 Late Cretaceous (Maastrichtian) palaeogeography and palaeokarst (Scotese et al, 1979).
Fig. 3.13 Early Palaeozoic palaeokarst and associated events and sequences. (The curve of global 3rd order cycles of relative change of coastal onlap has been derived from several sources.)
types of palaeokarst events. Some are related to single unconformities (SUK) recorded as depositional sequence boundaries (Vail's 3rd order cycles). Most of these are Type 1 sequence boundaries (SUK-1). However, there are also minor palaeokarst events associated with Type 2 sequence boundaries and to periods of subaerial exposure in shoaling-upward cycles (parasequences, punctuated aggradational cycles), herein grouped as PSUK events. Examples of these occur in the Pennsylvanian. PSUK palaeokarst reservoirs (Styles 1 and 5 of Encl. 3.1); they are characterized by thin profiles, good matrix and moldic porosity (Type A) and show good depositional control. SUK-1 events tend to show thicker profiles and a combination of matrix and moldic porosity with Types B and C aquifers.

A large number of palaeokarsts occur in relation to composite unconformities (CUK), commonly stacked at supersequence or megasequence boundaries. Stacked regressive cycles of sea-level generate optimum conditions for a potentially thick karst profile. These palaeokarst events are associated with long exposure times (more than 4 my.); they coincide with two, three or more 3rd order regressive cycles (represented as an arrow in the curves of coastal onlap) and have potential for thick palaeokarst profiles. CUK reservoirs correspond to Styles 2 to 4 (Encl. 3.1), with combination of Types B and C aquifers, typically with important structural control in the pattern of karst porosity. The best conditions for reservoir development and preservation are met when these regressive cycles correspond to important sea-level drops and intercalated transgressive cycles are relatively short lived. These are the CUK-A events; examples of them affected the Oligocene and the Chesterian (Middle Carboniferous). Less than optimum conditions occur when the overall regressive pattern does not involve large-scale sea-level falls and the transgressive and highstand parts of the cycle have long duration. These are the CUK-B events that commonly show an important quantity of karst porosity obliteration. CUK-B events occurred during the Norian-Hettangian and Aalenian.

CUK events are not evenly distributed through geologic time (Figs. 3.14 & 3.15). CUK-A events tend to occur during the evolution towards glacial periods (e.g., Devonian, Carboniferous, Cenozoic), coinciding with a generalized decrease in rifting activity and ridge volumes. These represent periods of palaeokarst occurrences toward higher latitudes. CUK-B events are common during the periods of decreasing glaciation and increasing rifting activity (e.g. Triassic and Jurassic).

3.5.3 Strategy for Unconformity Palaeokarst Plays

After the preliminary screening of global karst events and their palaeogeographic trends, strategy for unconformity-palaeokarst plays is outlined in the form of the following successive steps or phases: (1) recognition of unconformities and palaeokarst, (2) analysis, and (3) prediction of karst porosity.

3.5.3a Recognition of Unconformities and Palaeokarst

The precise recognition of an unconformity involves the integration/interaction of techniques of seismic interpretation, log correlation, biostratigraphy, petrology and geochemistry and development of a litho-chronostratigraphic framework with recognition of erosional vacuities and depositional hiatuses. In the identification of exposure facies it is important to avoid comparing rock fabrics without their sequential and regional context. Only very few rock fabrics by themselves can be considered as diagnostic of the different exposure facies and environments (Esteban and Klappa, 1983). Many of them can also occur as products of other sedimentary or diagenetic processes unrelated to the unconformity-palaeokarst environments. Any subaerially exposed carbonate formation is a potential candidate for the development of karst, but its identification requires careful documentation; for the interpretation of
Fig. 3.14 Distribution of global karst events compared with the 2nd order (supercycles) of global relative changes of coastal onlap and mid-oceanic ridge volumes (Vail et al, 1976).
Fig. 3.15 Distribution of global karst events compared with the 1st order cycles of sea level (Vail et al., 1976), percentage of continental flooding (Hallam, 1977), granite emplacement and major glaciations (Fisher, 1983).
karst implies an integrated drainage system of meteoric waters that includes conduit flow.

3.5.3b Unconformity-Palaeokarst Analysis

After its recognition, the unconformity-palaeokarst should be characterized as completely as possible in terms of a conceptual reconstruction of the environment (or succession of environments) during which the unconformity-palaeokarst originated. This reconstruction (unconformity-palaeokarst analysis) is not a superfluous academic exercise but an essential step to understanding and predicting porosity evolution in palaeokarst. It is at least as important as the sedimentological and environmental interpretations of the affected rock formations. However, the study of the exposure facies present some limitations. The thickness of the exposure profile and the intensity of the diagentic alteration are not proportional to the exposure time because of the influence of a large number of parameters (parent rock type, surface topography, palaeoclimate, palaeodrainage, hydraulic head, etc.). Also, because the strongly erosional nature of many unconformities, exposure facies and profiles can be reworked or destroyed; with that, the reconstruction of palaeoenvironments requires indirect methods. In essence, the unconformity-palaeokarst analysis involves consideration of the following aspects at regional and local scales:

1) Unconformity mapping. This is a subcrop map with representation of geomorphology of the exposure surface, with drainage patterns, local base levels, erosional lineaments, scarps, etc. Where one or several tectonic events occurred during or after exposure time, the reconstruction of the exposure surface prior to deformation should be attempted.

2) Exposure profiles. Surface and subsurface information on exposure profiles should be plotted on the subcrop map (soil and regolith types, thicknesses, karst facies, etc.).

3) Exposure environments. The interpretation of exposure environments should be superimposed on the basic subcrop map, with distinction of and detail of terrestrial and coastal areas and their zone of interaction, and the associated vegetation-microclimate zonation.

4) Post-unconformity events. Onlapping formations are displayed by plotting depositional patterns, tectonic features, onlapping geometries, etc., with special attention given to the basal contacts with the unconformity and the reworking of the exposure facies.

3.5.3c Karst Porosity Prediction

The prediction of trends and patterns of karst porosity requires consideration of the lithologies and porosity types existing at the time of onset of the subaerial exposure. Of particular interest is the presence of unstable mineralogies (aragonite, evaporite) in subaerially exposed formations and their degree of fracturing. The study of cement (pore-fill) stratigraphy is a useful approach that allows the distinction of the porosity destroyed or generated during burial of the unconformity and the transgressive onlap. In most cases, the key approach to the prediction of karst porosity is the analysis of the fracture pattern existing at the time of subaerial exposure.

The next step involves reconstruction of the drainage basin, with evaluation of the recharge area, amount of rainfall, base level, hydraulic head and preferential permeability networks controlled by structure and lithologies. Information on the karst facies and their stages of evolution provide useful constraints to reconstruction of the drainage model. In some cases, mostly in stable plateau karst areas, the surface palaeotopography gives an idea of the subsurface drainage patterns and location of the base levels. In general, surface drainage pattern are more ephemeral or rapidly changing than their subsurface counterparts. The ideal reconstruction of the drainage basin also include evaluation of allogenic recharge (rivers) and the contribution of
3.5.3d Integrated Model and Assessment

Ideally, the unconformity-palaeokarst analysis identifies the most satisfactory combinations of parent lithologies, palaeoclimate, exposure environments, geomorphology and drainage patterns for optimum reservoir development. However, in most cases this will be limited by the amount and quality of the available data. Actually, it is not possible to present a case history illustrating all aspects of the unconformity-palaeokarst analysis outlined above; the information normally available allows only partial completion of some of the steps and provide the basis for further refinement. This implies that porosity prediction involves hypothetical assumptions and indirect reasoning. It is within this context that the compilation of palaeokarst case histories in chapter 4 becomes very useful. They are useful references for analogies and trends when considering the relevance of the differences in lithologies, structural settings, timing, environmental conditions, stages of karst evolution, types of aquifers, reservoir styles and types of karst complexes.

3.6 ADDITIONAL EXPLORATION/PRODUCTION CONSIDERATIONS

Some additional general considerations are: (1) identification of karst breccias and fine-grained sediments, (2) predictability of reservoir parameters, (3) the origin of secondary chalky microporosity, (4) interpretation of surface topography: karst vs reef, and (5) source-rock constraints.

3.6.1 Identification of Karst Breccia and Fine-grained Sediments

The most common palaeokarst facies recognizable in subsurface samples are karstic breccias and fine-grained sediments. Karst breccias are matrix or clast-supported and monomictic or polymictic, with varying rounding; they occur in pockets, layers or pipes. Karst breccias may display chaotic assemblages of diverse composition, but some sequences are characterized by either local fining- or coarsening-upward sequences. Karst breccias form as surface mantling deposits (soils, regoliths or wedges flanking cliffs), sinkhole fill or karren morphologies. They also occur as cave sediments resulting from local collapse or after variable degrees of transportation from different sources. Karst breccias can be reworked in coastal cliffs and contain marine grains and organic borings. Strictly speaking, karst breccias are formed by either root activity in the soil zone or by cave collapse in late stages (vadose) of conduit evolution. Polygonally fitted breccias occurring in thin horizons or patches are possibly related to early stages of roof incision (collapse). However, thick intervals with fitted breccias (as well as some zebra fabrics common in MVT deposits) are probably related to hydrothermal processes and hydrofracturing in the deep phreatic zone. Different types of pressure-solution features are commonly present in karst breccias.

Breccias in caves (as cave sediments or as wall rock) can also be inherited from pre-existing tectonic or evaporite-collapse breccias (less commonly from slump deposits and debris flows in submarine or subaerial fans) rather than being directly formed by karst processes. These breccias of non-karstic origin found in the karst profile may be active aquifers and part of the karst system undergoing porosity enhancement or obliteration. These pre-existing breccias also enhance cave collapse and contribute to the resulting deposits.

In the interpretation of carbonate sections with breccias, tectonic, karstic or evaporite-collapse origin may not be mutually exclusive, and in many cases development is not simple. Pre-existing tectonic breccias can be karstified. Lithified cave sediments and wall rock are commonly fractured and brecciated and undergo another cycle of
karstification. Some tectonic brecciation also occurs after burial of the karst profile. Evaporite-collapse produced by meteoric-water solution is considered to be part of the karst processes. Horizons of evaporite-collapse breccias have developed preferential karst conduits in many reservoirs. On the other hand, karst breccias can be cemented by evaporites as part of the karst processes. In summary, it is commonly quite difficult to use a simple term, such as "tectonic" or "karstic" to characterize breccias in carbonate sections.

Fine-grained karst sediments are easy to recognize where they are brown-red (upper vadose zone) and/or where they are associated with typical speleothems and obvious caves. Very commonly, fine-grained palaeokarst sediments are carbonate-rich and locally dolomitic. Karst sediments are also dark gray, green or bluish (deep vadose and phreatic karst). Some occur in interstratified layers with variable amounts of grains of diverse origin (marine bioclasts, quartz, older fossils) and displaying diverse sedimentary structures (ripple, dune, cross-bedding, including bidirectional types, turbidites, slumps, etc.). As a result, fine-grained karst sediments, with or without sandy intercalations, have been erroneously regarded as part of the depositional history of the karstified formation. Where sediments from the onlapping sequence on a karst terrain infiltrate karst conduits, subsurface samples and well logs can be misinterpreted as depositional interfingering of two coeval lithologies. Petrographic hints of their karstic origin are provided by bizarre neomorphic microporoporitc fabrics that appear as concretions and palisades, locally with prismatic morphologies, or by intercalated carbonate cements, analytically determined to be consistent with speleothems. However, in many instances, elucidation requires extensive regional petrologic studies.

3.6.2 Predictability of Reservoir Parameters

Where palaeokarst is suspected or identified in the subsurface, the fundamental question is the prediction of the reservoir properties in terms of size (thickness and extent), trend and pattern of the permeability network and types and values of porosity and permeability. These parameters depend on the type of aquifers and palaeokarst complex, styles of reservoir, timing of the unconformity, regional evolution before, during and after exposure time and a series of local factors (structure, fracture patterns, size of the recharge area, hydraulic head, drainage pattern, lithologies, etc.). In essence, predictability of reservoir parameters requires the unconformity-palaeokarst analysis as outlined above. As a first approximation the parameters of karst reservoirs may be considered to be primarily controlled by depositional factors, local or regional structural features, or by a combination of the two (Fig. 3.7).

Depositionally controlled palaeokarst is associated with single unconformities (SUK events) and subaerial exposure surfaces in shallowing-upward cycles (PSUK events). It is characteristic of subaerial exposure of carbonate deposits shortly after deposition during exposure times of about 1 my or less. Reservoir parameters are controlled by depositional geometries, facies trends and lithologies. Karst profiles in PSUK events are relatively thin, except where related to pinnacle reefs in evaporitic basins and where stacked in multiple layers (Reservoir Style 5). Karst conduits are small and poorly integrated (excluding well developed reef framework porosity), and most of the drainage occurs as diffuse flow in primary intergranular, grain-moldic or micro-vuggy porosities. The solubility of the different carbonate minerals of the original deposit plays a major role in the karst evolution. Where aragonitic constituents (grains, cement, matrix) are present, leaching in the karst profile is very intense, but so is cementation (Harrison, 1975). In arid or semi-arid settings this alteration will be more intense in the phreatic zone. Where original constituents are of calcitic mineralogies, leaching and cementation occurs at a much reduced rate, and the karstified formation may retain large quantities of the original primary depositional porosity (James and
Karst reservoirs of SUK events show variable amounts of structural control.

Predominant structural control occurs in palaeokarst reservoirs developed under composite unconformities (CUK events) and, to some extent, below single unconformities (SUK events). The karst system developed during exposure times of more than 1 my., commonly between 4 and 40 my., but in some cases more than 200 my. Karst drainage is controlled by fracture patterns and structural dips; reservoir parameters tend to be highly heterogeneous. Karst profiles are up to several hundred meters thick, and they commonly contain well developed conduits. The mineralogy of the carbonate formation is essentially low-Mg calcite and dolomite. The different carbonate lithologies, thicknesses of beds and fabrics as well as the intercalated non-carbonate rocks (evaporites, shales, sandstones) do have an effect on the fracture patterns, and it is in this sense that some structurally-controlled karst reservoirs show some influence of depositional facies and patterns.

The large-scale structural control of karst development is also shown by the stratigraphic distribution of the corresponding unconformities. Outside these main cases of structural control, the stratigraphic range of the unconformity diminishes and the palaeokarst tends to be depositionally controlled. There is also a pattern of structural zonation in relation to orogenic belts. Most of the palaeokarst with MVT deposits and bauxites are essentially pre-orogenic and related to precursor movements. Bardossy (1982) noted a migration toward more external (outer) tectonic positions of younger karst horizons with bauxites. It appears that MVT karst occupies a more internal (inner) position than the bauxitic karst. Most cases of structurally controlled palaeokarst with hydrocarbons were developed in relation to post-orogenic unconformities, prior to post-orogenic subsidence and burial. The distribution of the different types and ages of palaeokarst in the Timan-Pechora basin is a good illustration of this general trend.

Karst reservoirs present numerous drilling and production problems; their prediction and control depends essentially on the type of karst reservoir. In general, drops in drilling bits and lost circulation are spectacular indicators of the presence of large conduits. However, absence of these type of accidents do not indicate absence of karst, only absence of large conduits in that part of the karst basin.

Horizontal drilling has dramatically improved the performance of some karst reservoirs (i.e., Rospo Mare, Italy; Dusser et al., 1988) by maximizing vadose karstic porosity development controlled by fractures. In general, similar results are to be expected in Mountain, Mediterranean and Hydrothermal Palaeokarst types. However, exploiting horizontal drilling should not be considered a universal receipt for all types of palaeokarst reservoirs. It is quite likely that horizontal drilling is of limited assistance in the cases where (1) the main reservoir is the zone of enhanced permeability in a Coastal Palaeokarst, (2) Plateau Palaeokarst without major deformation, and (3) Early Palaeokarst, depositionally controlled, associated with PSUK events.

3.7 DATABASE

A selection of 254 cases of palaeokarst is presented as a database with the stratigraphic parameters of related unconformity. From each palaeokarst region, the examples selected are those cases with different stratigraphic parameters in order to avoid both stratigraphic and geographic multiplicity. A large number of these case histories correspond to karstic bauxites, and are derived from Bardossy (1982). The database contains 50 cases of palaeokarst with hydrocarbon production. Only 30 representative cases of palaeokarst in outcrops and with MVT (Mississippi-Valley Type) deposits are included.

Figure 3.16 displays the case histories ranked by the youngest age (age
Fig. 3.16 Stratigraphic parameters of 254 case histories of palaeokarst ranked by the youngest age of onlapping unit. Arrows indicate concentration of cases of palaeokarst burial at the end of major tectonic events.
Fig. 3.17 Frequency distribution of palaeokarst case histories ranked according to the youngest age of the top of the onlapping unit. Time-gap interval = 50 my.

Fig. 3.18 Frequency distribution of palaeokarst case histories ranked according to the duration of the minimum gaps. Time-gap interval = 50 my.
Fig. 3.19 Frequency distribution of palaeokarst case histories ranked according to the duration of the maximum gaps. Time-gap interval = 50 my.

Fig. 3.20 Frequency distribution of palaeokarst case histories ranked according to the duration of the minimum gaps. Time-gap interval = 10 my.
Fig. 3.21 Frequency distribution of palaeokarst case histories ranked according to the duration of the maximum gaps. Time-gap interval = 10 my.
at the top) of the onlapping unit. There is a clear trend toward larger time gaps in connection with the Alpine orogenic cycle, which also corresponds to more than 40% of the total cases of palaeokarst in the data base. The shoulders (arrows) on the plot of the younger age of the onlapping units imply abundant cases in relation to the middle Oligocene CUK event, the Early Tertiary multiple CUK event, the Middle Cretaceous CUK and the Neo-Kimmerian CUK. The high-slope segments between the shoulders indicate periods of time when palaeokarst events were scarce or absent. There are fewer examples of older palaeokarst cases, and the shoulders and slopes for Paleozoic and Early Mesozoic data are not as well defined as for the Late Mesozoic and Tertiary. Figure 3.16 also shows that stratigraphic gaps for the Mesozoic cases are significantly smaller than those for the Tertiary or the Paleozoic cases; this reflects the relative tectonic tranquility that characterizes the Mesozoic.

The frequency distribution of palaeokarst case histories ranked according to the youngest age of the overlying unit (Fig. 3.17) displays the dominance of Late Cretaceous and Tertiary cases. A minor increase in the frequency of cases at 200-250 Ma and at 350-400 Ma reflects the burial of Late Hercynian and Caledonian palaeokarst structures. Figures 3.18 & 3.19) show predominance of case histories with time gaps of less than 50 my as well as decrease in frequency of cases with maxgaps over 250 my and mingaps over 200 my. These breaks in frequency are similar to those observed with time-gap intervals of 10 my (Figs. 3.20 & 3.21). Mingaps of less than 10 my are clearly dominant. It is very likely that a majority of cases within this group involve time gaps of 4-8 my.
CHAPTER 4
PALAEOKARST: CASE HISTORIES

by M. Esteban

4.1 TERTIARY/MESOZOIC UNCONFORMITY MEDITERRANEAN REGION

4.1.1 Introduction

The Tertiary/Mesozoic unconformity developed as a post-tectonic unconformity after the main phase of the Late Alpine orogeny. It is a multiple, composite unconformity formed by superposition or stacking of several Palaeogene and Miocene unconformities that truncate the fractured Mesozoic carbonate section at various depths. The Tertiary/Mesozoic unconformity shows particularly well preserved palaeokarsts in foreland areas overprinted by molasse basins, by post-orogenic sequences, or by late stage foredeep sediments and allochthonous, gravitational units. In these settings, the Mesozoic carbonate section is affected by rift and oblique tectonics with a variety of structures (e.g., vertical wrench faults, upthrusts, low angle thrusts, folds).

In the Mediterranean area the Tertiary/Mesozoic has excellent examples of palaeokarst with hydrocarbon production (Fig. 4.1). The Gulf of Valencia (offshore NE Spain) Casablanca, Montanazo, Amposta and Tarraco fields and good outcrops of the same palaeokarst event provide data for particularly good models. In Southern Italy, the Rospo Mare field is another example and similar palaeokarst settings are represented by the Nagylengyel field (SW Hungary) and Corbi Mari (Romania). In the Vienna basin the Tertiary/Mesozoic unconformity palaeokarst in the Triassic dolomites of the Calcareous Alps produces thick Lower Cretaceous and Upper and Lower Jurassic carbonates. Enhanced fracturing in massive dolomitic formations were conducive to the best karst development. Palaeogene sediments are commonly absent, but where present they occur as thin and discontinuous continental red beds and breccias. Thick and varied Palaeogene sedimentation occurred in subsiding troughs marginal to the uplifted areas, and there was synsedimentary tectonic deformation of the Palaeogene section along the contacts between uplifts and troughs.

3) Collapse. After late Oligocene and mostly during the Miocene there was...
Fig. 4.1 Major fields with production from the palaeokarst associated with the Tertiary/Mesozoic unconformity.
Post-orogenic, pre-Neogene palaeokarst reservoirs in the Triassic dolomites of the Calcareous Alps, overlain by Miocene formations, Matzen field, near Vienna, Vienna basin (Ladwein, 1988).
progressive collapse and foundering of the uplifted areas. The karstified Mesozoic is successively onlapped by transitional and marine Miocene sequences. Karst systems continued to operate in the remaining structural highs, but the pre-existing karst porosity tended to be obliterated. Block faulting was very important, locally accompanied by oblique and gravitational tectonics. There was continued nappe movement in the foredeep areas. As uplifted areas collapsed, some parts of the Palaeogene troughs were uplifted and karstified.

4) Subsidence. During late Miocene and Pliocene generalized subsidence affected the productive fields and neighboring areas, with the final onlap of the remaining highs. During this subsidence phase there is peak maturation and migration of hydrocarbons into the available karst reservoirs. Most of the source rocks are in the lower Neogene (or upper Oligocene) section that were deposited in a variety of lacustrine and transitional marine environments. Apparently, Jurassic and Triassic source rocks formed in some areas.

4.1.2 Stages of Cyclic Karst Development

Modern karst analogs and the Palaeogene and Neogene stratigraphic record offer the basis for a model of karst development in different stages according to the relative sea-level positions for each depositional sequence (Fig.4.3). Development of karst will be different during the low stand, the transgressive phase and the high stand.

Low-stand karst: It is characterized by a large recharge area and a high hydraulic head, with potential for very intense karstification (Fig.4.4). During the Neogene (and probably Palaeogene as well) the low stand corresponds to a climatic minimum with abundant rainfall and vegetation cover, which also contribute to the maximum potential for generation of karst porosity. Coastal mixing corrosion involving karst waters with an excess of pCO2 is likely to occur on the flanks of the structures and may involve the flanking wedge deposits. The low stand stage typically occurs immediately after a tectonic pulse, with uplift. This feature, combined with the intensity of the surface karst processes, produces a juvenile landscape with important erosion rates and abundant deposition of clastic carbonates (breccias, conglomerates, sands and silts) on the flanks of the structure. The instability of the surface morphologies is conducive to common olistostromic deposition, particularly in the area affected by coastal processes.

Transgressive stage karst: A rise in relative sea-level produces a reduction of the recharge area and hydraulic head with the corresponding diminution in karst activity (Fig.4.5). Part of the karst conduits created during the low stand become incorporated in the zone of flow stagnation and porosity obliteration. Transgressive sediments onlap and rework the karstified structure; typically these sediments are skeletal carbonates rich in red algae, benthic forams, mollusks, bryozoans, etc., with variable amounts of mixing with clastic carbonates and without hermatypic coral reefs. The old karst surface and mantling deposits (regoliths, soils, scree, etc.) represent reworking in coastal-marine environments (beaches). Headward erosion, with a base level determined by the encroaching sea, is the dominant process. Coastal mixing corrosion is possible, but it mostly will affect the transgressive sedimentary section.

High-stand karst: During a high stand of relative sea-level the recharge area and the hydraulic head of the karst system will be reduced to a minimum; most of the old karstified structure will be in the area of flow stagnation and porosity obliteration (Fig.4.6). The high stand of sea-level is considered also to correspond to relatively dry and hot climates resulting in reduced recharge of aggressive waters. Karst landscape rapidly evolves into senility. Deposits of clastic carbonates are minor and finer grained. Miocene high stands (and some of the Palaeogene) are characterized by coral-reef complexes prograding over the transgressive and low-stand sedimentary sections. During the Pliocene and the Quaternary there is no development of coral reefs in the Mediterranean, and the
Fig. 4.3  Schematic structural cross-section of the Tarragona basin, depicting the different karst events recorded in the Mesozoic section.
Coastal karst aquifers in the Catalan Ranges. A. Hydrogeologic cross-section, Vandellós area, just north of the Ebro delta (Pascual et al, 1986). B. General karst model for the Mesozoic of Gulf of Valencia, based on present-day hydrology and applicable to phases of sea level low-stand. PS= Piezometric surface, HW= High water.
Fig. 4.5  Karst model during a transgressing sea-level.  PS = Piezometric surface.

Fig. 4.6  Karst model during a sea-level high-stand.  PS = Piezometric surface.
high stand carbonate formations are similar to the transgressive carbonates.

4.1.3 Casablanca Field Case History

BASIN/LOCATION
Tarragona basin of the Gulf of Valencia, NE Spain, offshore part of the Alpine Catalan Ranges in the area off the Ebro delta, provinces of Tarragona and Castellon (Fig. 4.7).

BASIN TYPE
Bally: 221 - Ramp with buried grabens.
Klemme: IIIA - Continental rifted basin, craton and accreted zone rift.

GEOLOGICAL CONTEXT
Major unconformity and subaerial exposure severely eroded Mesozoic carbonates during the Senonian to Oligocene. Most of the area probably was already emergent in the late Albian and was subaerially exposed until early Miocene.

STRUCTURE
Late Oligocene-Burdigalian rifting resulted in complex NE-SW trending grabens. Predominant block faulting occurred in Langhian-Serravallian (Figs.4.8 & 9), but distribution of lower Miocene facies demonstrates early graben formation. Important modifications occurred in the Messinian and Plio-Quaternary. Late Neogene volcanics occur in the SE part of the Gulf of Valencia. Block faulting in the Gulf of Valencia had important horizontal (strike-slip) component, a fact ignored by some workers. To some extent, Tertiary faults are reactivated Mesozoic and late Hercynian structural trends.

The Casablanca field is the largest of the Castellon high fields (Montanazo, Castellon, Salmonete, Tarraco and Angula). Other fields (Fig.4.7) occur on the Amposta high and the Tarragona Platform (Amposta, Dorada).

STRATIGRAPHY
The Jurassic and Cretaceous are a thick succession (up to 5500 m thick in the onshore, but normally much less) of shallow-marine limestones and dolomites with important lateral variations in facies and thicknesses. On some structural highs there are subcrops of Triassic continental red beds and restricted marine carbonates and Palaeozoic metamorphic rocks. Most lower-middle Miocene units were originally included in the Alcanar Group, consisting of the Alcanar Conglomerate, the Amposta Limestone, and Tarraco Shale. These represent a variety of continental, shallow-marine and deeper water environments. Modern stratigraphic usage restricts the term "Alcanar" to the basal depositional sequence and distinguishes the Casablanca-Sant Carles-Salou units as a series of poorly defined depositional sequences with a wide variety of depositional facies. The Castellon Group is composed of sands, silts and shales, forming a thick, prograding shelf wedge.

RESERVOIR
Depth: Top average 2600 m; spill point at 2733 m (Fig.4.7).
Lithology: Upper Jurassic shallow-marine limestone and dolomite with vuggy, cavern and karst-enlarged fracture porosity and a well developed karst profile. Locally, reservoir is a heterogeneous breccia.
Thickness: Effective reservoir is 90-152 m thick below the unconformity, but evidence of karst occurs at deeper levels.
Porosity: Average of 7-12%, locally up to 28%; core samples show only 1-3%. Most porosity developed by karstification during the Early Tertiary uplift, with some hydrothermal leaching during Late Neogene.
Distribution: Karst porosity occurs in the upper 200 m of the Mesozoic formations, but it presents very irregular lateral distribution. Up to three cave levels can be distinguished, with the possibility of independent oil/water contacts. In the more mature karst profiles, the shallow karst facies show intense porosity destruction by karst sedimentation and collapse. Karst breccias grade laterally into scree and alluvial deposits.
Reservoir Style: (Esteban-Wilson classification) 3a. Linear trends associated with rift blocks.

SEAL
Marls and shales of onlapping Miocene Cambrils and Castellon groups.
General cross-section of the Catalan Coastal Range and the Gulf of Valencia, NE Spain (Bayó 1987, unpublished course notes). Paleokarst reservoirs in the fields were produced by composite unconformities in the Paleogene, modified and buried in the Neogene. Modern karst in the Catalan Coastal Range inherits and modifies Cenozoic paleokarst products.
Fig. 4.8 Generalized structural map and cross-section of the Casablanca field, Tarragona basin, Gulf of Valencia (García-Siñeriz et al., 1980).
Fig. 4.9  Seismic line. Casablanca field area, showing faulted, fractured karstic Mesozoic carbonates below Tertiary shale.
FIELD
Discovery: 1975.
Approximate Number of Wells: 100
(in Gulf of Valencia).
Area: 11 x 2.5 km (Casablanca field).
Estimated Ultimate Recovery: 90
million bbl for the Casablanca field only;
satellite fields in the Castellon High are
estimated at 60 million bbl. The overall
Gulf of Valencia is estimated at 270
million bbl.
Oil/Gas Characteristics: Gravity 33.7°
API, 0.2% sulfur, gas/oil ratio of
155 cf/bbl, temperature 150°C.
Drive Mechanism: Fresh-water drive
from onshore.

SOURCE
Burdigalian Casablanca marls and shales,
dysaerobic marine deeper water facies.
Onlapping unconformity and palaeokarst
reservoir and well developed in the
graben areas (i.e., Tarragona trough). Migration in the Pliocene.

COMMENT
First accounts considered the reservoir to
be of a fracture type. Karst facies are
difficult to define in most logs and seismic
lines.

4.2 PRE-PENNYSYLVIANIAN KARST
NORTH AMERICA

4.2.1 The Significant
Mississippian-Pennsylvanian
Unconformity in Western
North America

Of the several regional
unconformities developed during the
Palaeozoic across the North American
craton (Sloss, 1963), three show common
gigakarst features such as sinkholes,
cavern, collapse breccias, enlarged joints,
joint fillings, and terra rosa. These
unconformities are: (1) the post-Sauk
sequence (pre-Middle Ordovician)
surface; (2) the complex of pre-intra-
Devonian unconformities which actually
are Middle Silurian to Middle Devonian
and/or pre-Frasnian black shale; and (3)
the complex unconformity which post-
dates the Madison and is pre-Chesteran
and/or pre-Pennsylvanian. The last
erosional period began toward the end of
the Meramecian; it is marked in the
northern basins (Williston, Michigan,
Windsor) by regressive evaporite
deposition. Because latest Mississippian
(Chesteran) deposits are absent over the
Transcontinental arch, commonly a basal
Pennsylvanian sand, or red bed sequence,
overlies karstic Madison Group
carbonates, which span most of the
Mississippian. The complex
unconformity above the Madison is well
documented at the edges of the Williston
basin, over the Transcontinental arch
(Figs.4.10 & 11), including the area of the
Wyoming shelf, across Ancestral Rockies,
and throughout the Southwest from the
Grand Canyon to New Mexico. Solution
along fractures and attendant breccias
and collapse zones laid the groundwork
for much lead-zinc-ba rite mineralization
in the Rockies; also several important oil
and gas fields are developed in reservoirs
associated with this surface: for example,
(1) at the Madison edgeline in
Saskatchewan (Coleville-Buffalo Coulee),
(2) Elk Basin, in the Big Horn basin of
Wyoming, and (3) Lisbon in the Paradox
basin. Several fields and outcrop
occurrences of the unconformity are
discussed below. Outcrops are plotted on
Figure 4.10, which is a pre-
Pennsylvania palaeogeologic map,
showing character of karstic features at
the Mississippian-Pennsylvanian
unconformity.

Oil fields in Saskatchewan and
north-central North Dakota are made up
of eroded knobs and cherty detritus
produced by palaeodrainage patterns
cutting through cuestas along the Lower
Mississippian subcrop edge. Here the
group of partly cemented and solution-
altered Mississippian carbonates were
buried by early Mesozoic red beds and
evaporites. In Montana and western
North Dakota, on the west side of the
Williston basin, Charles evaporites at the
top of the Madison are overlain by the Big
Snowy Group of Chesteran age (possibly
with slight disconformity). Channel
sandstones and red beds in the Lower to
Middle Pennsylvanian Tyler Formation,
just below the Amsden Formation overlie
the top of the Big Snowy Group. Farther
south on the Wyoming shelf, a well
known large oil field, Elk Basin, with
more than 85 million bbls of oil, exists at
Fig. 4.10 Pre-Pennsylvanian palaeogeologic map showing Mississippian strata surrounding eroded highlands of Ozark dome, Transcontinental arch, Nemaha ridge and N-S and NWW trends of southwestern USA (McKee and Crosby, 1975).
Fig. 4.11  Palaeogeography of the USA during Middle Mississippian time (late Osagean and early Meramecian, about 345-353 Ma) (Sando, 1988).

Fig. 4.12  Location map of Elk Basin field in Big Horn basin, with structure on top of Tensleep Formation (Pennsylvanian) (McCaleb and Wayhan, 1969).
the northern end of the Big Horn basin. It was discovered, in 1946, that production could be developed in this old field at the top of the thoroughly karstified Madison (Mission Canyon beds of Osagian age). The Madison is overlain here by Amsden Pennsylvanian strata. Cavern collapse and evaporite solution breccias 200-300 ft into the Madison are well documented (McCaleb and Wayhan, 1969). Reservoir performance is highly complex because of multiple blocks of collapsed strata below the unconformity.

The unconformity above the Madison of the Wyoming shelf has been studied intensively by Sando (1988) who notes that the upper 400 ft of the unit has been affected by karstic erosion with westward drainage toward the Cordilleran miogeocline; this drainage is well marked by isopachs of the overlying basal transgressive Darwin Sandstone Member of the Amsden Formation. Several major river systems cut across a low plain with about 200 ft of average relief. Southeast Wyoming, an upland area, was part of the Transcontinental arch, which was uplifted increasingly along the Pennsylvanian Front Range of the Ancestral Rockies. Across the Wyoming shelf major evaporite units, carbonate bedding surfaces, and fault blocks in the Madison formed conduits for water drainage.

The block faulted pattern in the Madison (Leadville) also resulted in extreme karstification across the axis of the Transcontinental arch in Colorado (DeVoto, 1988), particularly across the Front Range, Sawatch, and Uncompaghre uplifts of the Ancestral Rockies (Fig.4.10). Deeply incised paleovalleys, up to 200 m deep in places, formed on the Leadville (Madison) surface with sinkholes and caverns even affecting older Palaeozoic strata. These valleys with an attendant northeast joint pattern have controlled distribution of extensive mineralization.

The Paradox basin to the southwest of the Transcontinental arch was downwarped in the Laramide orogeny but it was a Carboniferous basin which contains northwest-trending major Pennsylvanian fault blocks on which the ever-present Madison karst was developed (Baars, 1966). The Lisbon field is developed at the edge of one of these blocks in karst-modified Leadville (Madison) reservoir rocks (Miller, 1985). Ultimate production is estimated to reach 43 million barrels of oil and 250 BCF of sour gas. The reservoir is in dolomite of multiple origins, being in part syngenetic associated with now-dissolved evaporites as well as that formed by later ingress of meteoric water which extensively dolomitized the micritic sediment. A small amount of hydrothermal karst-associated, coarse sparry dolomite is also present. Most of the dolomite is associated with the pre-Molas (Pennsylvanian) unconformity and late-phase leaching and collapse brecciation of the carbonate. As much as 100 ft of Leadville has been eroded on the crest of the Lisbon structure. Silicification, fracturing, and brecciation are common features of the reservoir, and anhydrite solution breccias are also important.

Even farther south, in southern New Mexico, the major unconformity is pre-Pennsylvanian and is developed at the top of the Lake Valley Group (Madison equivalent). This area is along the crest of the southern terminus of the Transcontinental arch; no Chesteran or Morrow is present. The hiatus is some 35 million years, as in Wyoming. The effect of this unconformity on the petrography of the underlying Lake Valley sediments has been studied by Meyers (1974, 1978, 1988). The surface is marked by relief of up to a few tens of meters. Meteoric water flow below the unconformity is evidenced by zoned phreatic cements around crinoidal debris, etching of these cements, filling of residual pore space by micrite and microspar (eluviated lime mud) and clay and detrital quartz silt, presence of microbreccia, anastomosing veins and fissures filled with detritus, and beds of chert breccia (Meyers, 1974; 1978; 1988). The karst surface shows development over low-relief terrain, mainly in the vadose zone, on only partly cemented crinoidal hash, and with chemically active ground water flow in the western and northwestern parts of New Mexico. Erosion may have begun during Late Mississippian time, but most activity was in the Early Pennsylvanian.

Summary of features known to indicate karst development at the post-Madison (Leadville and Lake Valley)
surface over western North America from Saskatchewan to New Mexico:

2. Enlarged solution joints filled with quartz sand.
3. Caverns generally with collapse breccias.
4. Sinkholes generally with collapse breccias.
5. Mappable regional drainage patterns.
6. Incised palaeovalleys with hundreds of feet relief.
7. Bedded evaporite solution breccias in beds below unconformity.
8. Cherty detrital breccias or conglomerates, residual sediment - in places reservoir rock.
10. Meteoric phreatic zone cements with zonation and etching traceable below unconformity, eluviated lime mud deposits as micrite-microspar.
11. Extensive dolomitization associated with the unconformity. Dolomitization of open marine micritic facies.
12. Lead-zinc mineralization and sparry dolomite associated with hydrothermal karst.

4.2.2 Elk Basin Field Case History

BASIN/LOCATION
Near NE edge of Tertiary Big Horn basin, on the Wyoming shelf as it developed during Carboniferous time. Elk Basin is a NNW-trending Laramide anticline on the flank of the Big Horn uplift (Fig. 4.12).

BASIN TYPE
Bally: 222 - Foredeep dominated by block faulting.
Klemme: IIA - cratonic margin, composite, foredeep.

Big Horn basin is a crustal sag, foreland composite basin of a Late Cretaceous-Tertiary basin developed over a Palaeozoic shelf area.

GEOLOGICAL CONTEXT
Megakarst development in upper part of Madison limestone-dolomite section below unconformable top forms carbonate breccia-dolomite reservoir with secondary porosity; entrapment is structural. Internal reservoir performance controlled by collapse blocks of Madison (Fig. 4.13).

STRUCTURE
Asymmetric anticline, trending NNW with a thrust fault on the steep northeast flank (Fig. 4.14).

STRATIGRAPHY
Madison is of Mission Canyon (Mississippian) age at the top. The whole formation is about 800 ft thick of mostly shallow, subtidal carbonate with alternating restricted marine and high-energy oolitic and crinoidal hash. It is rather thoroughly dolomitized. Mostly grainy beds are present at top, which is erosional. It overlies the Cottonwood Canyon, a shaley carbonate which disconformably overlies Devonian carbonates. Overlying Amsden is thick over the collapse zone in the uppermost of the Madison, occurring along the anticlinal structural high. Collapse zone affects 200-350 ft of Madison. Five subdivisions are recognised in the collapse zone (Figs. 4.15 - 4.18). All of A (upper) zone has so many collapse blocks that poor reservoir continuity, and hence prediction, made field development very difficult. Some traceable beds of collapse-solution breccia are present throughout the field, but they are not believed to be evaporite solution breccias.

RESERVOIR
Depth: 5000-6500 ft.
Lithology: The Madison is divided into four major units A-D. A zone (with five subdivisions) generally is characterized by dolomite, solution breccia, with dense argillaceous zones. A1 is sugary dolomite, developed from a calcarenite above a karst breccia zone. A1b is medium-grained sugary dolomite with patchy porosity followed below by a dense shaly dolomite zone. A2 and A3 are also sugary dolomite; the subjacent unit is a widely distributed major solution breccia. B zone is finer grained sugary dolomite. C and D
Fig. 4.13 Isopach map of Pennsylvanian Amsden Formation showing post-Madison topography in Elk Basin field (McCaleb and Wayhan, 1969).

Fig. 4.14 Structural contour map on top Madison Formation, Elk Basin field. Cross-section in Figures 5.14.11-12. Numbered wells in Figures 5.14.10-13 (McCaleb and Wayhan, 1969).
Fig. 4.15 Typical Elk Basin Madison well log, showing zonation. Location is shown on Figure 5.14.9 (McCaleb and Wayhan, 1969).
Fig. 4.16 North-south stratigraphic cross-section.

Fig. 4.17 Idealized general stratigraphic cross-section showing karst solution zones in Madison reservoir.
zones are also finer grained, plugged by dolomite and anhydrite (Figs. 4.15 - 4.18). Thickness: 780 ft for gross thickness. Net pay within all zones totals 188 ft. Distribution: Madison is widespread over western USA. Unconformity on top of Madison is also widespread in the Rocky Mountain area. Trap and production are controlled by local structural configuration. Environment of Deposition: Open marine to restricted marine. Some tidal flat-sabkha at top; i.e., evaporites, but these are minor. Porosity and Permeability: Porosity 10-13.5%; Permeability 3-368 mD. Permeability in A zones is much better, averaging >65 mD. Reservoir Style: (Esteban-Wilson Classification) 3b. Linear trend caused by later formed anticline which transects a megakarst at the unconformable top of a massive dolomite unit. SEALS Overlying Amsden shale is top seal. Oil-saturated, fine-grained impermeable dolomite occurs below reservoir but there is oil-water contact.

FIELD DETAILS

COMMENTS
1. Apparent similarity of detailed karst sequence (A-B solution breccia) with widely traceable middle zone of Ellenburger sequence in Emma Field in Texas. This Madison zone might also represent a collapsed cavern passage.

2. Problem of distinguishing such strata from evaporite-solution breccias whose continuity is also widespread.

EXPLORATION CONCEPTS
The field is a combination of a clearly defined structural trap associated with a specific type of diagenesis, notably cavern collapse at major unconformity associated with development of megakarst (Figs. 4.17 & 4.18).

4.2.3 Oklahoma City Field Case History

BASIN/LOCATION
Central Oklahoma in Oklahoma County on a north-south ridge (a very narrow horst block).

BASIN TYPE
Bally: 221 - Foredeep adjacent to A subduction zone (or in this case adjacent to wrench or megashear system).
Klemme: IIA - Continental multicycle, craton margin, composite.

The field is located on the northern shelf of the Anadarko basin aulacogen. This is a neutral shelf area between the Anadarko basin and the McAlester foreland basin.

GEOLOGICAL CONTEXT
Karstified Arbuckle limestone-dolomite was affected by at least two unconformities developed along the crest of the narrow Nemaha Ridge. Coincidence of the well-developed karst reservoir and the structural trap resulted in the field.

STRUCTURE
The field reservoir is found in a pronounced faulted anticlinal fold located at the southern end of the Late Palaeozoic Nemaha Ridge, which is a long and narrow horst trending SSW for more than 200 miles across eastern Kansas and central Oklahoma. Multiple periods of structural growth occurred along the ridge in the Palaeozoic; the most pronounced uplift is pre-Pennsylvanian.
Fig. 4.18 Correlation cross-section of upper karstic part of Madison. Wells located in Figure 5.14.9 (McCaleb and Wayham, 1969).
The east-bounding fault of the anticline trends NNW and has a maximum throw of 2400 ft. The producing closure of the structure is 1200 ft.

STRATIGRAPHY
The classical Midcontinent, (Arbuckle Mountains) section is present in the field (Fig. 4.19). The lowest productive zone is the Arbuckle dolomite. It is overlain by Pennsylvanian strata at the crest of the structure where doming and erosion has removed about 300 ft of upper Arbuckle. In sequence above the Arbuckle (3300 ft thick) lies the rest of the Ordovician, about 1200 ft of the Simpson Group, Viola Limestone and Sylvan Shale, and the Hunton Group (Siluro-Devonian) about 300 ft thick with a major unconformity between it and the complex Pennsylvanian shale, sandstone, and limestone section.

RESERVOIR
Depth: 6400 ft.
Lithology: The lowest and original productive horizon was Arbuckle Dolomite, which has been altered by meteoric water. It is fine crystalline dolomite with solution cavities, vugs, and fractures. Bit drops of several feet were common indicating holes in the reservoir as much as 7 ft in diameter. Zavoico (1929) noted an additional unconformity-derived sediment, detrital reworked Simpson sands, above the pre-Pennsylvanian unconformity.
Thickness: Oil column of 700 ft. The most porous and cavernous zone of Arbuckle is 300 ft thick, from 200-250 ft to 500 ft within the formation. Net pay is commonly only 200-170 ft.
Distribution: The unconformity at the base of the Pennsylvanian is widespread in the southwestern USA but the field is structurally limited.
Environment of Deposition: Tidal-flat environment. Dolomite of West Spring Creek Formation is laminated and stromatolitic.
Porosity and Permeability: Porosity and permeability have little if any relationship to depositional fabric. Some intergranular porosity, most is secondary, vuggy, cavernous, fracture porosity.
Reservoir Style: (Esteban-Wilson Classification) 3a-3b. Linear trend related to anticlinal structures fold and fault. Note that "detrital zone" (reworked Simpson sand lenses) exhibit Style 4 (wedges above unconformity) (Fig. 4.19).

SEALS
Pennsylvanian shales which blanket the faulted structure form both lateral and vertical seals.

FIELD DETAILS
Discovery Date: December, 1928.
Method of Discovery: Surface anticline drilled first. Then subsurface geology studied for field development. Later substantiation was by seismic data.
Number of Wells: 112 Arbuckle wells from a total of 1800 wells.
Area: Whole field, 13,770 acres. Arbuckle acreage 2460-about 18% of total field; 12 x 4.5 mi.
Estimated Ultimate Recovery: 18,200,000 bbls. of oil, 68 BCF of gas.
Oil Characteristics: Very high gravity oil. 38-41 ° API.
Temperature: 72-110° F. for the oil.
Pressure: Very high pressures. 2686 psi at 6400 ft.
Drive Mechanism: Water drive. Was early flooded "coned in" with water because of excessive production by city lot spaced wells. Only 24% of oil in place was recovered.

COMMENTS
Interesting history of exploitation in which the rapid production of Arbuckle oil and the practice of intermittent shut down and then flush production resulted in water "coning" which by-passed most of the oil. The very permeable "honeycomb" Arbuckle karstic carbonate permitted very easy and rapid ingress of water.

EXPLORATION CONCEPTS
An example of how multiple unconformities prepare a reservoir which is structurally controlled.

4.3 PENNSYLVANIAN-LOWER PERMIAN KARST OF SOUTHWESTERN U.S.A.

The rapid changes of sea level and ensuing disconformities, so frequent
Fig. 4.19 Geologic cross-sections through Oklahoma City field, showing relation of detrital zone and "Siliceous lime" (Arbuckle) (Zavoico, 1929).
during the Pennsylvanian, gave opportunity for high relief that was affected significantly by incursion of meteoric water. The abundant carbonate buildups were particularly susceptible to such meteoric diagenesis. Typical Pennsylvanian "reefs" or "mud mounds" are composed of internally fractured and brecciated micrite replete with phylloid (platy) algae, many with a codiacean-like cortical structure. Presumably these originally were of aragonite and lightly calcified, and when subjected to fresh water soon after deposition, they were easily leached, causing good porosity and permeability. The algae apparently accumulated as detritus within lime mud (few, if any, growth or framework clusters are present); the accumulations were subjected to early marine cement in the form of druse fans and fringes. In Early and Middle Pennsylvanian the sponge-like Chaeretes helped to form reef-like buildups, and a tiny branched stromatoporoid, Komia, added to the detritus. In Late Pennsylvanian and Early Permian time, masses of encrusting tubular foraminifera and another tiny sponge-like encruster, Tubiphytes, were able to bind and stabilize lime mud and organic debris and thereby form ecologic reefs. Thus, cores of all of these buildups commonly consist of small binding and encrusting forms with little organic framework. More than half the volume of the buildups may consist of cross-bedded flanks of debris which generally does not have as good reservoir properties as the core itself. The buildups are commonly capped by grainstones which may be tightly cemented and furnish a seal.

Individual buildups are typically small, a few tens of feet thick, a few hundreds of feet across, and a half to a mile long. For this reason individual reservoirs may be small, but because the buildups are often composite, superimposed and overlapping, multiple oil-water contacts are common in many fields which may produce from large composite carbonate masses. Furthermore, the leached algal plate fabric or algal plate grainstone has great permeability.

Although major unconformities with well developed megakarst are practically unknown within Pennsylvanian-Wolfcampian strata, meteoric diagenesis is so common in these buildups that literally hundreds of small fields are known in southwestern USA. These represent Styles 5b and 2b of the Esteban-Wilson classification of karst-meteoric reservoirs. The following table 4.1 shows fields that have been documented fairly well:

Several of the above fields are in clusters across a Strawn platform on the eastern shelf of the Midland basin. These are individual mounds, oriented approximately NE, formed during the Desmoinesian (Strawn) to Missourian (Canyon) transgression. They grew below the Missourian shelf margin and include both Strawn- and Canyon-aged strata (Desmoinesian to Missourian). The fields include North Knox, Claytonville, Rowan, Hope, Hope NW, Esteban, Stone, Lake Trammel, Nena Lucia, I.A.B., Jameson, Millican, Higgins Ranch, H.J. Strawn, Neva West, and Hulldale.

Strawn and Canyon buildups produce widely over other shelves in West Texas and in the Paradox basin, where the major Aneth field eclipses other production. Its ultimate recovery is estimated at greater than 300 million bbls and 300 BCF gas. Late Pennsylvanian and Wolfcampian buildups are well known on the northern shelf of the Delaware basin, both as shelf margin rims (Townsend-Kemnitz) and as "patch reefs" on or adjacent to N-S fault trends within the shelf itself (e.g. Saunders and Anderson Ranch). Furthermore, outcrops of Pennsylvanian and Wolfcampian strata in Utah, New Mexico, Colorado, and West Texas contain well developed and well-exposed buildups identical in age and facies to those producing in the adjacent basins. These have been studied in detail and are described in many publications (e.g. Wilson, 1975).

Two known major mound-reef complexes are giant fields:

1. Scurry mass or Horseshoe Atoll whose cumulative reserves are more than 2.54 billion bbls.

2. Scurry mass or Horseshoe Atoll whose cumulative reserves are more than 2.54 billion bbls.

<table>
<thead>
<tr>
<th>FIELD</th>
<th>COUNTY</th>
<th>STATE</th>
<th>AGE OF RESERVOIR</th>
<th>UNIT OF RESERVOIR</th>
<th>SIZE (m bbls)</th>
<th>REFERENCE</th>
</tr>
</thead>
</table>

Table 4.1 Fields with documented Pennsylvanian-Wolfcampian meteoric diagenesis.
Schatzinger, R.A., 1983, Phylloid algal and sponge-bryozoan mound to basin transition: A Late Palaeozoic facies tract from the Kelly-Snyder field, West Texas: Harris, P.M. (ed.), Carbonate Buildups: SEPM Core Workshop no.4, p. 244-303.

2. Indian Basin field near Carlsbad, New Mexico, which is one of the USA's major gas fields, several TCF reserve.

4.4 MIocene Karst in Coral Reefs, Far East

Miocene carbonate platforms and buildups were very abundant and well developed in different regions of the Far East. Most of the oil found in carbonates occurs in buildups. Platforms tend to be reservoirs for gas only (Kayoun et al., 1981). Platforms are regionally extensive blankets (up to about 200 m thick) with a wide variety of skeletal carbonates including coral shallow water patch reefs. Buildups are laterally discontinuous accumulations of skeletal carbonates (locally up to more than 1 km thick), and are called pinnacle reefs when present in a conical-ellipsoidal shape. Extensive flat-top buildups are called banks or platform-buildups. Figure 4.20 shows characteristic seismic expression of Miocene buildups. The literature on the carbonates of the Far East considers these buildups as the result of coral reef growth catching up with rising sea level in areas surrounded by relatively deep waters. However, many of the drilled buildups fail to show evidence of coral reef framework or even significant amounts of coral debris. The best coral reef frameworks in the area are interpreted as small patch reefs in shallow parts of extensive carbonate platforms (Jordan and Abdullah, 1988), but are very scarce in buildups or pinnacle reefs, where most of the cored samples show carbonate slope facies with variable amounts of mixing with lagoonal or shallow water (non-coral reef) sediments. For convenience, we refer to these buildups as pinnacle reefs with the understanding that their depositional geometries can be strongly modified by subsequent karst erosion processes.

Present-day karstification on outcrop of Miocene carbonates produces spectacular tower-karst landforms that mimic, at least in part, reef morphology.

The Miocene carbonates of the Far East present several (2-4) stages or cycles of development separated by discontinuities in reef growth with episodic subaerial exposure and karstification followed by renewed reef growth or by sedimentation of deeper water marine carbonates and marls. The biostratigraphy of these stages or cycles is poorly known; this report assumes their correlation with the global karst events (Late Aquitanian, Langhian, Late Serravallian and Late Tortonian-Messinian).

Karst processes occurred immediately after deposition and were particularly intense in reef rocks modifying primary (intergranular, framework) porosities and occurred during or soon after mineralogical stabilization (calcitization) of the original aragonite and high-Mg constituents. Most of the productive intervals show combinations of vuggy and moldic porosities with secondary chalky microporosity. Karst enhanced fracture porosity was also an important contributor. Loss of circulation and large caverns have been encountered in some wells (N. Sumatra), but are not an abundant type of karst in these Miocene carbonates. Some fields (Nido, Philippines) show evidence of confined karst aquifers with active circulation during burial after compaction.
Fig. 4.20 Seismic sections A, B, and C, across typical pinnacle and platform buildups in Luconia (Doust, 1981).
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Jordan, C.F., Jr., and Abdullah, M. 1988. Lithofacies analysis of the Arun reservoir, North Sumatra, Indonesia; in Giant Oil and Gas Fields, a Core Workshop. *SEPM Core Workshop No. 12*, p. 89-117.


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### Styles of Unconformity Reservoirs and Associated Paleokarst

<table>
<thead>
<tr>
<th>UNCONFORMITY TYPE</th>
<th>POTENTIAL DIAGENETIC RESERVOIR TYPES</th>
<th>EXAMPLES</th>
<th>COMMENTS</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. <strong>BLANKETS OVER BROAD ARCHES</strong></td>
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<tr>
<td>Reservoir seal</td>
<td></td>
<td>Hunton, Oklahoma</td>
<td>Low paleotopography, wide &amp; thin over regional arches, low structural complexity.</td>
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<td></td>
<td></td>
<td>Fahud, Natih, Um ad Dalkh, Middle East</td>
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<td></td>
<td></td>
<td>Luling, Texas</td>
<td></td>
</tr>
<tr>
<td>2. <strong>MOSAICS OF REMNANT HIGHS</strong></td>
<td></td>
<td>Yates, Texas (in part)</td>
<td>Non-linear pattern that can be purely depositional or purely erosional, commonly a combination of both. Pattern can be related to diapiric domes or to intensively and repeatedly deformed areas.</td>
</tr>
<tr>
<td>a. Depositional-erosional</td>
<td></td>
<td>Elk Basin, Wyoming (in part), Golden Lane, Mexico (in part)</td>
<td></td>
</tr>
<tr>
<td>b. Stacks of convergent unconformities</td>
<td></td>
<td>Parkman, Saskatchewan</td>
<td></td>
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<tr>
<td>3. <strong>LINEAR TRENDS</strong></td>
<td></td>
<td>Pennsylvaniaian algal mounds</td>
<td>Barely emergent mounds, shoals or patch reefs with local, convergent unconformities in particular case of 5b.</td>
</tr>
<tr>
<td>a. Fault</td>
<td></td>
<td>Baugh, New Mexico</td>
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<tr>
<td>b. Anticline</td>
<td></td>
<td>Fateh, Middle East (in part)</td>
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<tr>
<td>4. <strong>WEDGES AND MANTLES ABOVE UNCONFORMITY</strong></td>
<td></td>
<td>Albion-Scipio, Michigan</td>
<td>Linear structure developed before, during or after unconformity. Combinations of 4a &amp; b are common. Where thermokarst is important, reservoirs can be preferentially developed in lows.</td>
</tr>
<tr>
<td>a. Depositional-erosional</td>
<td></td>
<td>Casablanca, Spain, Renqiu, China</td>
<td>Potentially deep reservoirs and multiple cave levels.</td>
</tr>
<tr>
<td>5. <strong>LAYERS OF SHALLOWING CYCLES</strong></td>
<td></td>
<td>Poza Rica, Mexico</td>
<td>Consequent or subsequent sediments related to residual deposits or to onlapping sequence. Reservoir could be entirely formed by submarine exposure process in some settings. Commonly associated with flanks of 2 &amp; 3.</td>
</tr>
<tr>
<td>a. Layer-cake</td>
<td></td>
<td>Wolfcamp, Gunnnx, Midland Basin, Texas</td>
<td>This includes debris flow breccia, chert residues and regoliths, coastal deposits, fans, etc. with variable quantities of silicilastic deposits. Characteristics of sealable karst.</td>
</tr>
<tr>
<td>b. Layer-mound</td>
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<td>Casablanca, Spain (in part)</td>
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<td></td>
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<td>Murbank, Thamama, Abu Dhabi</td>
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<td>Smackover, Jay, Walker Creek, Texas</td>
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<td>Qatar, Berri, Ghawar, Saudi Arabia</td>
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<td>Coulommes, Sunniland, Florida</td>
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<td></td>
<td></td>
<td>Michigan Basin Pinnacle Reefs</td>
<td>Multiple shallowing upwards hemicycles (parasequences) with intermittent short exposure during or shortly after deposition. Also as stacked carbonate mounds or buildups, grading to type 2b with low growth rates.</td>
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<tr>
<td></td>
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<td>Scurry County, Fairway James Lst, Texas</td>
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<td>Intisar, Libya</td>
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<tr>
<td></td>
<td></td>
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<td>Multiple and thin reservoir horizons.</td>
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</table>

**Encl. 3.1** Styles of reservoirs and traps at unconformities and associated reservoirs.