Maturity of kerogen, petroleum generation and the application of fossils and organic matter for paleotemperature measurements

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Cover Picture: Middle ordovician conodont element.

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YASER ABDULLAH ABDULLAH AL-MASHRAMAH

Abstract: The origin, composition and maturity of kerogen, and its conversion to petroleum is reviewed. Various methods for estimating the maximum temperature to which kerogen and potential source rocks were subjected are also reviewed. Several kerogen maturation parameters can be distinguished and grouped into two categories: chemical paleothermometres and biological paleothermometres. Biological paleothermometres are largely based on the colour and optical properties of kerogen and fossils with an organic composition or with an organic component in their mineralized skeleton. The consistent changes in colour and reflectance through thermal gradients (burial depth) have made vitrinite, conodonts, graptolites, scolecodonts, acritarchs, spores and pollen, invaluable for maturation studies and prediction of hydrocarbon windows.

Spores, pollen and vitrinite are frequently used as thermal indicators in lacustrine and near-shore marine strata ranging in age from the Devonian to Recent. Spores and pollen can be used to assess the thermal alteration at low temperatures, whereas vitrinite can be used to estimate the degree of maturation of organic matter at low to high temperatures, even to the graphite-grade metamorphic facies. Conodonts are most common in carbonate facies and used as thermal indicators of marine strata ranging in age from the Cambrian to the uppermost Triassic They can provide information about thermal alteration at temperatures as high as 500 °C, a temperature far above the destruction of unicellular palynomorphs. Graptolites are most common in fine-grained clastics of Ordovician and Silurian age, i.e. in sedimentary rocks in which pollen, spores and vitrinite are virtually absent. Scolecodonts are common in Ordovician–Devonian shallow marine carbonates and show optical properties similar to those of graptolites and vitrinite.

Key words: Kerogen, thermal maturation, thermal alteration, petroleum generation, paleothermometres, vitrinite, conodonts, graptolites, palynomorphs.

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Termal mognad av organiskt material, oljebildning och tillämpning av fossil för paleotemperaturbestämning

YASER ABDULLAH ABDULLAH AL-MASHRAMAH

Sammanfattning: Det här examensarbetet ger en översikt över organiskt material (kerogen) i Jordens yngre lagrade berggrund och hur detta material omvandlats till olja och naturgas genom olika geologiska processer. De viktigaste är tryck, temperatur och tid. Med fortskridande sedimentpålagring förs det organiskt rika sedimenten successivt ned på större djup, där de genomgår en så kallad mognadsprocess till följd av ökad temperatur och tryck. Vid mognadsprocessen sönderdelas de stora molekylerna i materialet och omvandlas gradvis till olja och naturgas. Vid tryck- och temperaturökningen förändras sedimentens struktur, t.ex. med avseende på porositet, packning och mineralsammansättning, och även färg och glans på fossil som består av organiskt material. Inom oljeprospektering analyserar man därför ett flertal av dessa fossil (biologiska termala indikatorer) för att uppskatta begravningsdjup och relativ temperatur. I denna studie redovisas särskilt hur mikrofossil förändras som ett resultat av ökad temperatur. Sporer och pollen är viktiga termala indikatorer i sediment som avsatts i sjöar, medan conodonter, scolecodonter och graptoliter är viktiga för sediment avsatta i havsmiljöer. Conodonter är vanliga i kalkstenar från ordovicium till trias och är centrala för detta arbete. Dessa små tandelement genomgår en färgförändring från ljust gulaktig vid låga temperaturer till svarta vid höga temperaturer. Denna förändring beskrivs i en skala från 1-5, där fem indikerar en temperatur på mer än 300 °C. Optimala betingelser för oljebildning, det så kallade *oljefönstret*, är mellan 60 och 120 °C, motsvarande ca 1,5-2,0 på skalan. I detta arbete ingår även ett praktiskt moment, där conodonter utsattes för en temperatur av 500 °C under 24 timmar. Färgförändringarna dokumenterades genom fotografering efter förutbestämda tidsintervall. Experimentet lyckades inte reproducera tidigare publicerade resultat angående conodonters färgförändring vid ökande temperatur. Troliga orsaker diskuteras kortfattat.

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1 Introduction

Commercial accumulations of oil and gas are restricted to sedimentary basins in which organic matter, primarily kerogen, has been converted to liquid and gaseous hydrocarbons. With increasing temperature and burial depth, the organic matter alters progressively, mainly because hydrogen and oxygen are lost in excess to carbon. This thesis focuses on petroleum generation and the origin, composition and maturity of kerogen, with particular emphasis on biological palaeothermometers, such as vitrinite, pollen, spores, acritarchs, graptolites, scolecodonts, and conodonts. The consistent colour changes through thermal gradients have made vitrinite, organic-walled fossils and fossils with an inorganic component in their mineralized skeleton invaluable for prediction of hydrocarbon windows.

The aim of the thesis is to review how kerogen is converted to petroleum and the various methods for measuring the maximum temperature to which potential source rocks were subjected. The aim is also to compare the main indicators of thermal maturity zones and the main zones of hydrocarbon generation.

2 Kerogen

Kerogen is generally defined as disseminated organic matter in sediments that is insoluble in normal petroleum solvents (Selley 1985). It is a mixture of organic compounds (long-chain biopolymers) that contain carbon, hydrogen and oxygen with minor amounts of nitrogen and sulphur. The term is applied to organic matter in oil shales that yield oil upon heating and is regarded as the prime source for petroleum generation (Durand 1980). Thus, it is used for organic matter that converts to petroleum (crude oil and natural gases) after burial and heating in sedimentary basins. Kerogen is distinguished from bitumen because it is insoluble in normal petroleum solvents whereas bitumen is soluble (Selley 1985). Kerogen has four sources: lacustrine, marine, terrestrial, and recycled. It is not in equilibrium with the surrounding liquids (e.g. Tissot & Welte 1978; Selley 1985). Most oil has been formed from lacustrine and marine kerogen. Terrestrial organic matter generates coal and the cycle is inert (e.g. Tissot & Welte 1978; Selley 1985; Hunt 1995).

In petroleum studies the kerogen are classified into three basic types (I, II, and III) based on the ratio between their C, H, and O content, (e.g. Tissot & Welte 1978; Selley 1985). Types I and II are referred to as sapropelic kerogen, and type III is known as humic kerogen. Sapropelic kerogen is formed through decomposition and polymerization products of fatty, lipid organic material, such as algae and spores, and is used to designate finely disseminated kerogen, exhibiting an amorphous structure after destruction by acids (e.g. Tissot & Welte 1978; Selley 1985; Hunt 1995). Humic kerogen is largely produced from the lignin of higher land plants. Types I and II predominantly generate oil, type III primarily generates gas and some waxy oil. Type I kerogen is of algal origin and mostly consists of lipids with small amounts of fats or waxes. It is characterized by a high proportion of hydrogen (high H/C) relative to oxygen (low C/O) (e.g. Tissot & Welte 1978; Fig. 1). The polyaromatic nuclei and heteroatomic ketone is less important. Lipids represent a high proportion because it is generated due to a selective accumulation of algal material and due to a severe biodegradation of the organic matter during deposition (e.g. Tissot & Welte 1978). Type II, referred to as liptinic kerogen, is common in petroleum source rocks and oil shale. It has a high H/C ratio (1.28) and low O/C ratio (0.1). It is derived from algal detritus in marine environments, but also contains a mixture of phytoplankton and zooplankton (Selley 1985). The polyaromatic nuclei and heteroatomic ketone is more important in this type than in type I (e.g. Tissot & Welte 1978). Type III has a relatively low H/C ratio (less than 1 or equal to 0.84) and a high C/O ratio (equal to 0.2). This type includes a high proportion of polyarmatic nuclei and heteroatomic ketone without ester groups and is low in aliphatic compounds. Type III is derived from woody plants and is less favourable to oil generation than other types. It generates abundant gas at greater depth (e.g. Tissot & Welte 1978). The chemistry of the different the kinds of kerogen are shown in Table 1.

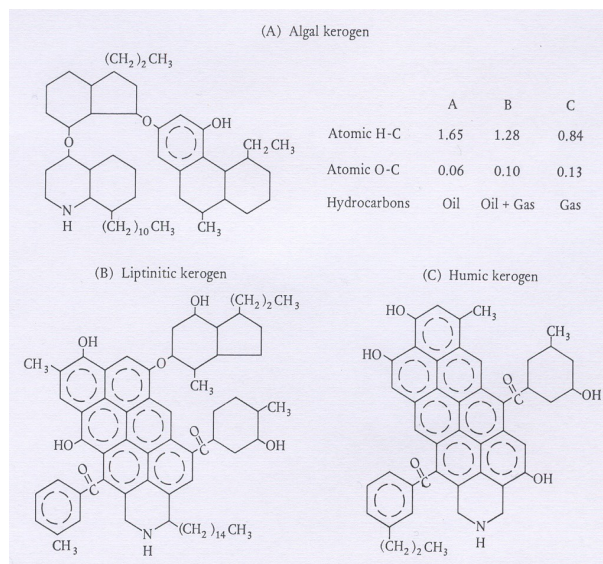*Fig.1.* Molecular structure of different types of kerogen. From Selley (1985, fig. 5.9).

3 Generation of petroleum
3.1 Introduction

The generation of petroleum by kerogen maturation depends on a combination of temperature, as a function of the depth of burial, and time. Kerogen is mostly formed in shallow subsurface environments. With increasing burial depth in a steadily subsiding basin, the

Table 1. H-C and O-C ratios in different types of kerogen. Modified after Selley (1985, table 5.4).

	H/C	O/C	Petroleum type
Type I, algal	1.65	0.06	Oil
Type II, liptinic	1.28	0.1	Oil and gas
Type III, humic	0.84	0.13–02	Gas

kerogen is affected by increased temperature and pressure. After burial and preservation, organic matter can go through three phases leading to kerogen degradation:

Diagenesis. – This phase occurs in shallow subsurface environments at low temperatures and near-normal pressures. It includes two processes, biogenic decay supported by bacteria, and abiogenic reactions (Selley 1985). Diagenesis results in a decrease of oxygen and a correlative increase of the carbon content. It is also characterized by a decrease in the H/O and O/C ratios (e.g. Tissot & Welte 1978).

Catagenesis. – This phase is marked by an increase in temperature and pressure, and occurs in deeper subsurface environments. It results in a decrease of the hydrogen content due to generation of hydrocarbons. Petroleum is released from kerogen during this stage. Oil is released during the initial phase of the catagenesis, at temperatures between 60 and 120 ºC. With increasing temperature and pressure (approximately 120–225 ºC), wet gas and subsequently dry gas are released along with increasing amounts of methane (e.g. Tissot & Welte 1978; Selley 1985). Catagenesis is characterized by a reduction of aliphatic bands due to a desubstitution on aromatic nuclei with increased aromatization of naphthenic rings.

Metagenesis. – This is the last stage in the thermal alteration of organic matter. It occurs at high pressures and temperatures (200 to 250 ºC) in subsurface environments leading to metamorphism and a decline of the hydrogen-carbon ratio. Generally only methane is released until only a carbon-rich solid residue is left (Fig. 2). At temperatures over 225 ºC, the kerogen is inert and only small amounts of carbon remain as graphite (Selley 1985; Fig 2). The vitrinite reflectance is used to indicate metagenesis because this stage leads to rearrangement of the aromatic nuclei (e.g. Tissot & Welte 1978).

The temperature that leads to maturation of kerogen can be estimated by many techniques, such as the colour and reflectance of organic matter in the rocks, measuring of the carbon ratio, analysis of clay mineral diagenesis, and fluorescence microscopy (Selley 1985; Hunt 1995). The colour of kerogen is dependent on the degree of maturation and its chemical composition and structure (Selley 1985). Pollen and spores are for instance originally colourless. With increasing burial depth and temperature they change from light to dark (Hunt 1995). One commonly used system is the Spore Colour Index of Barnard et al. (1976) and the Thermal Alteration Index developed by Staplin (1969). The reflectance of vitrinite is widely used for determining the maturation of organic matter in sedimentary basins and the maturity of potential source rocks. The reflectance is measured optically and only on vitrinite group macerals, since other macerals mature at different rates (Hunt 1995). The vitrinite macerals are separated from the sample by solution of hydrofluoric and hydrochloric acids (Selley 1985).

There are two hypothesis describing how petroleum is generated, an organic and an inorganic. The inorganic or abiogenic hypothesis postulates that oil forms by reduction of carbon or its oxidized form at elevated temperatures deep in the Earth (Selley 1985; Hunt 1995). Thus, methane can form through various types of metamorphic and igneous processes, and is found trapped in some minerals (Craig et al. 2001, p. 159). Although there is some evidence for an abiogenic origin of some methane, overwhelming geochemical and geological evidence shows that most petroleum is formed from accumulation of organic matter trapped and altered in sedimentary rocks (Hunt 1995).

According to the organic theory, the origin of the petroleum follows two pathways from living organisms and goes through three stages (diagenesis, catagenesis and metagenesis). The first pathway represents petroleum formed directly from hydrocarbons synthesized by living organisms or from their molecules (Hunt 1995). This pathway involves an accumulation of hydrocarbons formed by dead organism and other hydrocarbons formed by bacterial activity and chemical reactions at low temperatures (Hunt 1995). The second pathway involves thermal maturation of converted organic matter, such as lipids, proteins and carbohydrates, into kerogen during diagenesis (Selley 1985; Hunt 1995). With increasing burial depth and temperature, the organic matter progressively cracks to bitumen and liquid petroleum (Selley 1985; Hunt 1995).

3.2 Oil generation

As pointed out above, oil is mainly formed during the catagenetic phase, which is also known as the oil window or oil zone. With increasing temperature more molecular bonds are broken (H/C or O/C) and hydrocarbon molecules and aliphatic chains are formed from the kerogen. The hydrocarbons generated are C_{15} to C_{30} biogenic molecules, which have low to medium molecule weight (e.g. Tissot & Welte 1978). When burial and temperature increase, light hydrocarbons are generated due to cracking, and increase the proportion of source rock hydrocarbons and petroleum (e.g. Tissot & Welte 1978; Hunt 1995). Subsequently, the hy-
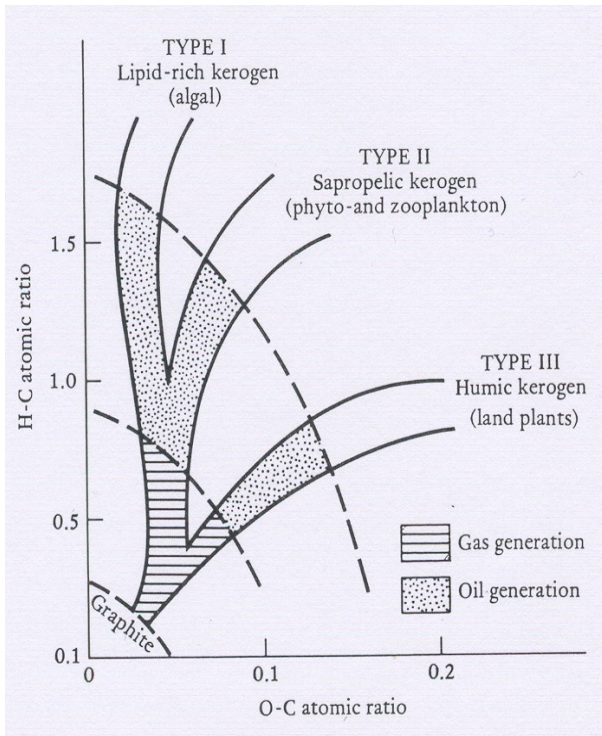
Fig. 2. Diagram showing the maturation pathway of the three principal types of kerogen. From Selley (1985, fig. 5.10)

Fig. 3. Oil generation window offshore South Padre Island, Gulf of Mexico, showing an increase in generation of hydrocarbons at a depth of 2.5 km and a peak at 4 km. From Hunt (1995, fig. 5.8)

Fig. 4. Oil generation window in the Green River Formation of the Uinta Basin, Utah, USA, showing an increase in generation of oil at 2.6 km and a peak at a present-day depth of 3.0 km. From Hunt (1995, fig. 5.11).

drocarbons convert to wet gas with an increased amount of methane. The temperature in which oil is generated and expelled from the source rock ranges from 60 to 160 ℃. This temperature interval is known as the oil window (Hunt 1995)

Figure 3 shows the range of oil windows in the South Padre Island, Texas. The oil window corresponds to the depth interval 3.0 to 4.9 km. At 2.5 km there is an exponential increase in the C_6–C_7 hydrocarbons generated and there is a peak at about 4.0 km (Hunt 1984, 1995). The stratigraphic interval above and below the oil window is referred to as immature and postmature for oil generation, respectively (Hunt 1995).

The lacustrine black shale facies of the Green River Formation in the Uinta Basin, Utah, contains type I of kerogen, whereas types II and III dominate in the alluvial facies of the formation (Hunt 1995). At around 2.6 km there is an exponential increase in the yield of C_5–C_7 hydrocarbons (Anders & Gerrild 1984; Hunt 1995; Fig. 4). The yield reaches its maximum at a present-day depth of 3.0 km, equivalent to 95 ℃. At 3.7 km there is a marked decrease in the generation of C_5–C_7 hydrocarbons. It must be emphasised, however, that the basin has been uplifted about 1.8 km (Sweeney et al. 1987), so that the true oil peak-yield temperature seems to be close to 140 ℃ (Hunt 1995). Thus, the oil window in the Green River Formation is at a lower depth than that in South Padre, but, as noted above, this is because of uplifting of the Uinta Basin.

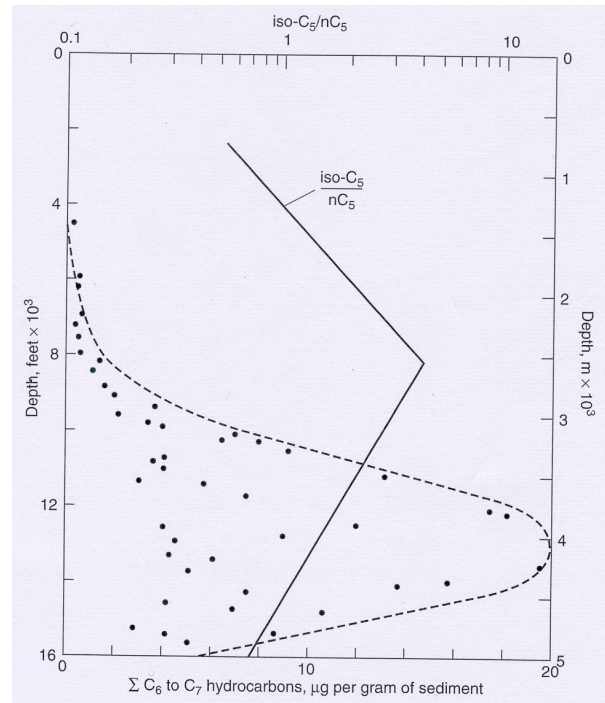The oil window and the correlation between temperature and subsurface depth are also well known in the Aquitaine Basin of France (Le Tran 1972; Hunt 1995). The oil yield peaks at 100 ℃, equivalent to a depth of 3.0 km. The Aquitaine Basin is of particular

Fig. 5. Oil and gas windows in the Aquitaine Basin of France, showing a peak of oil generation at 3.0 km and the beginning of substantial gas generation within the oil window. From Hunt (1995, fig. 5.14).

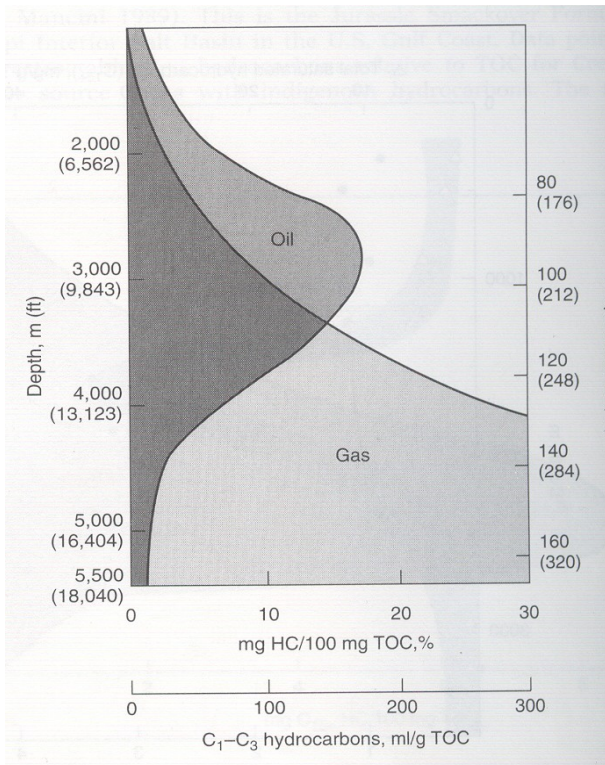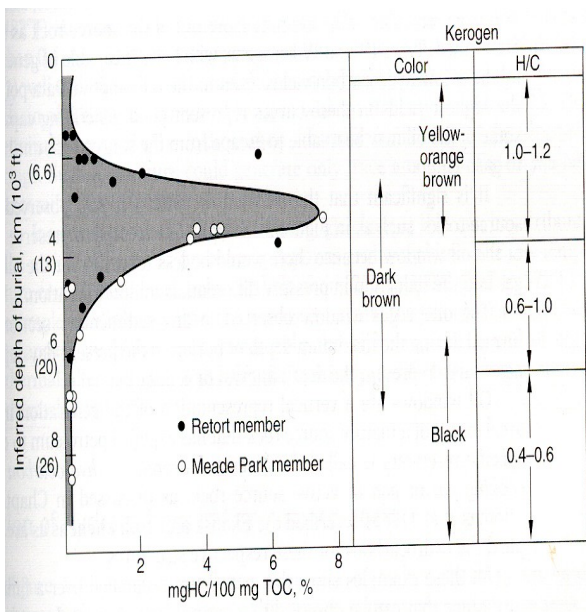
Fig. 6. Oil generation window in the Permian Phosphoria Formation, western United States, the oil window and changes in kerogen composition (H/C) and colour with burial depth. From Hunt (1995, fig. 5.15).

interest because the beginning of substantial gas generation occurs within the oil window (Fig. 5). The gas window peaks at 150 ºC, equivalent to a depth of approximately 5.0 km (Le Tran 1972; Hunt 1995).

As pointed out above, the kerogen structure changes during the three phases of maturation. It must be emphasized, however, that the oil-yield curve and its relation to subsurface temperatures can change from basin to basin. During the changes, the hydrogen/carbon ratio of the kerogen decreases because the kerogen provides hydrogen to form the oil (Hunt 1995). This change causes a change in the colour of the kerogen, from yellow-orange brown to dark brown and finally black (Claypool et al. 1978; Hunt 1995). Figure 6 shows the oil windows and the gradual changes in H/C ratio in the Permian Phosphoria Formation, western United States. The peak for C_{15+} hydrocarbons occurred at a depth of 3.3 km (Claypool et al 1978). The oil window gradually starts to disappear at a depth of 5 km.

These studies show that the zone of intense oil generation and the peak for oil-yield differ from basin to basin because the oil generation is influenced by several factors, such as migration of oil out of the source rock and conversion of oil to gas (Hunt 1995). When the generation of oil is more than the migration out of the source rock + the conversion of oil, the yield increases, but when the migration and conversion of oil to gas is greater than generation of oil, the yield will decrease (Hunt 1995).

3.3 Gas generation

Natural gas is the gaseous phase of petroleum. It typically contains methane (a high percentage), ethane together with other hydrocarbons up through the nonanes (C_9H_{20}), and nonhydrocarbons, such as carbon dioxide, hydrogen, nitrogen, argon, helium, and hydrogen sulfide (e.g. Selley 1985; Tissot & Welte 1978; Hunt 1995). Liquid hydrocarbons dissolved in the gas can also be present (e.g. Tissot & Welte 1978). Methane is always a major constituent of gas. Natural gas has multiple sources, both organic and inorganic. Biogenic gas, for instance methane, carbon dioxide and hydrogen sulphide, may be formed by bacterial activity, in particular by methanogenic bacteria, at the surface or in very shallow subsurface environments (e.g. Tissot & Welte 1978; Hunt 1995). Carbon dioxide is also formed during diagenesis as part of the elimination of oxygen from the kerogen. Carbon dioxide can, however, also be generated by degradation of crude oil during catagenesis (e.g. Tissot & Welte 1978).

At greater depths and temperatures, the thermal cracking of kerogen and coal generates gas, in particular methane. Generation of methane is normally subordinate to heavier hydrocarbons during early catagenesis. During late catagenesis and metagenesis, hydrogen sulphide and substantial amounts of methane are generated from kerogen (primary cracking; Hunt 1995). The cracking of oil to methane and light hydrocarbons is called secondary cracking (Hunt 1995).

Wet gas, i.e. a condensate consisting of liquid hydrocarbons dissolved in the gas, is essentially formed during the catagenetic stage. As noted above,

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methane (dry gas) can be formed bacterially during early diagenesis, but most of the dry gas in the subsurface was formed thermally during catagenesis and metagenesis (e.g. Tissot & Welte 1978).

Some gas components can have an inorganic origin. Nonhydrocarbon gases, such as of CO_2, H_2S, H_2 and N_2, may be derived from volcanic activity or they can have geothermal sources (e.g. Tissot & Welte 1978). There is also evidence for an abiogenic origin of some methane (Selley 1985). Inert gases, in particular helium and argon, are produced by radioactivity. Helium is, for instance, produced by disintegration of radioisotopes from the uranium and thorium families, and argon is derived from disintegration of potassium (e.g. Tissot & Welte 1978; Hunt 1995).

4 Thermal maturation indicators

In oil exploration it is of prime importance to determine the maximum temperature to which the kerogen or a source rock has been subjected. Thus, it is necessary to have paleothermometres. There are several kerogen maturation parameters. The various techniques that can be used can be grouped into two categories: chemical paleothermometres and biological paleothermometres (Selley 1985).

The chemical methods include studies of organic and inorganic matter. The most important techniques for studying the maturity of organic matter are the carbon ratio method, the electron spin resonance (ESR) technique, and pyrolysis. Inorganic chemical paleothermometres include clay mineral analysis and studies of fluid inclusions (Selley 1985). Both organic and inorganic indicators are important for reconstruction of thermal histories of sedimentary basins.

The carbon ratio give an estimate of the paleotemperature because there is a gradual decrease in the ratio of organic carbon and an increase in the amount of fixed carbon during maturation of kerogen (Selley 1985). The electron spin resonance method is based on the number and distribution of free electrons in the atomic fraction of kerogen. With increasing maturity the electron characteristics change (Selley 1985). Pyrolysis, i.e. heating of kerogen, provides immediate information on maturity and source rock potential. It is based on the fact that hydrocarbons are expelled from the kerogen with increasing temperature, and hydrocarbons expelled at different temperature intervals are determined in order to compare the amount of petroleum already generated and the total amount capable of being generated (Selley 1985; Hunt 1995).

With increasing temperatures, argillaceous rocks are subjected to various diagenetic reactions and changes. Studies of clay mineral reactions have increased during the past decades along with intensified petroleum exploration, because clay minerals undergo various diagenetic changes during burial. These clay minerals include in particular smectites (montmorillonites), illite and kaolinite. With increasing depth, pressure and temperature, the clay will re-crystalize and smectites will progressively change to illite (Selley 1985). Smectites generally disappear at temperatures over 95 ºC (Dunoyer de Segonzac 1970; Olsson-Borell 2003). Thus, the change in the chemical structure of clay from smectites to illites indicates the required temperature for oil generation. The kaolinite is also destroyed during diagenesis, but the disappearance of kaolinite is not indicative of a certain temperature or burial depth (Olsson-Borell 2003). As temperatures increase to the upper limits of oil and gas generation and incipient metamorphic processes, kaolin and illite will recrystalize into mica (muscovite) or chlorite (Selley 1985; Olsson-Borell 2003).

Biological paleothermometres are largely based on the colour and the optical properties, for instance the light reflectance and fluorescence, of organic matter (kerogen) and fossils, such as pollen, spores, acritarchs, graptolites, scolecodonts, and conodonts. These fossils have an organic composition or have an organic component in a mineralized skeleton (conodonts). With increasing temperature and burial depth, their organic matter alters progressively, mainly because hydrogen and oxygen are lost in excess to carbon (e.g. Marshall 1990), and the consistent colour changes through thermal gradients have made microfossils invaluable for prediction of hydrocarbon windows.

5 Biological paleothermometres
5.1. Conodonts
5.1.1 Introduction

Conodonts are marine tooth-like microfossils considered to belong to an extinct group of eel-like fishes. They range from the upper Cambrian to the uppermost Triassic, and have become the premier microfossils for dating shallow marine carbonates (e.g. Armstrong & Brasier 2005). They had a global distribution and can be found in a wide range of marine environments. Conodonts are most common and diverse in limestones, particularly in those representing tropical nearshore environments (e.g. Armstrong & Brasier 2005; Milsom & Rigby 2010). Deeper water conodont assemblages are generally less diverse. The majority of the conodonts show some degree of facies dependence, which suggests that the conodont animals lived close to the sea floor (Armstrong & Brasier 2005). The abundance of conodonts varies due to depositional conditions. When the deposition is rapid conodonts become rare, perhaps only 1–2 specimens per kg of rock, and sometimes conodonts are absent. On the contrary, a low rate of sedimentation can result in a high abundance of conodonts (up to several thousands per kg; Epstein et al. 1977).

The individual conodont elements consist of calcium phosphate (calcium carbonate fluorapatite) with organic carbon trapped in their crystal structure (e.g. Symons & Cioppa 2002). The most detailed study of the chemical composition is by Pietzner et al. (1968), who concluded that the formula for the mineral matter is as follows: $Ca_5\,Na_{0.14}\,(PO_4)_{3.01}\,(CO_3)_{0.16}\,F_{0.73}$

($H_2O)_{0.85}$). Conodont elements resemble small fish teeth and typically range from 0.1 to 5 mm in size. The elements were arranged in a feeding apparatus different to the jaws of modern animals (Marshall et al. 1999). The apparatuses were usually formed of seven to eight pairs of elements.

Conodonts are firmly classified within the phylum Chordata (e.g. Armstrong & Brasier 2005). Sweet (1988) subdivided the conodonts into two classes, Cavidonti and Conodonti. The class Cavidonti includes two groups, Proconodontida and Belodellida, which range from the upper Cambrian into the Devonian and are distinguished by thin-walled and smooth-surfaced elements (Sweet 1988). The class Conodonti comprises five groups: Protopanderodontida, Panderodontida, Prionodontida, Prioniodinida, and Ozarkodinida (Sweet 1988).

True conodonts (euconodonts) appeared during the late Cambrian and became extinct in the late Triassic. They reached a maximum diversity during the Early and Middle Ordovician with a diversity decline during the Silurian (Armstrong & Brasier 2005; Milson & Rigby 2010). Conodonts radiated again during the Devonian and were still abundant during the Carboniferous. Following a general decline during the Permian, they disappeared during the late Triassic mass extinction (e.g. Armstrong & Brasier 2005).

Conodonts are important biostratigraphic indicators in the Palaeozoic, especially for dating shallow marine carbonates, and are also widely applied in palaeoecological and palaeobiogeographical analyses (e.g. Sweet 1988; Armstrong & Brasier 2005). The first appearance datum (FAD) of selected conodonts has also been used for defining several of the stages within the Palaeozoic Erathem (e.g. Gradstein et al. 2004) and for defining the base of the Triassic (e.g. Yin et al. 2001). Conodonts are especially used as stratigraphic tools in shallow marine carbonate successions where graptolites are absent, and can be used together with several other types of microfossils, e.g. acritarchs and spores. They are also applied in studies of metamorphism, geothermometry, and structural geology, and for assessing the oil and gas potential of sedimentary basins (Epstein et al. 1977).

Conodonts are important paleothermometres because their colour changes with temperature (e.g. Epstein et al. 1977). Thus, they are frequently used as indicators of thermal alteration of sedimentary rocks. The change in colour of conodonts is known as conodont Colour Alteration Index (CAI), which is used in determining the oil and gas windows (Epstein et al. 1977). The colour alteration of conodonts has been known for over 100 years (Harley 1891), who noted that irons compounds caused the change in the colour (Bustin et al. 1992). Since the 1970s, use of conodonts as palaeotemperature tools has rapidly become a stan-

Fig. 7. Pale yellow conodonts (CAI<1) from the Ordovician Kope Formation, Kentucky, USA. From Epstein et al. (1977, fig. 2).
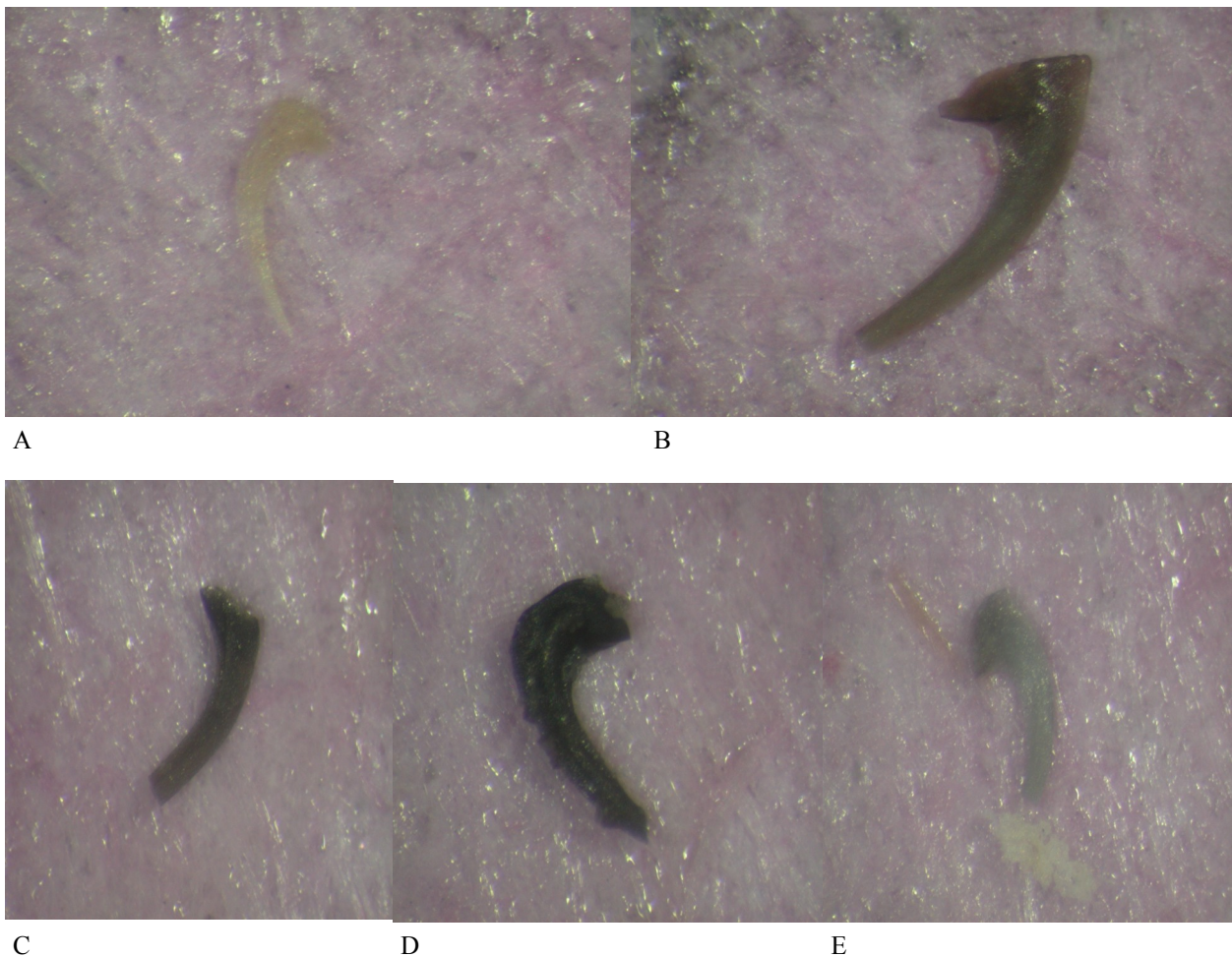
Fig. 8. Ordovician conodonts from Estonia (A), Great Britain (B), Jämtland, Sweden (C) and Västergötland, Sweden (D and E) showing different CAI values. A, CAI=1; B, CAI=2; C, CAI=3; D, CAI=5; E, CAI=6–6.5.

dard practice in exploration for hydrocarbons in the Palaeozoic and Triassic.

5.1.2 Colour Alteration Index (CAI)

As pointed out above, conodonts are composed of calcium phosphate and organic matter. The organic matter undergoes chemical changes, which are related to depth of burial and duration of burial (e.g. Symons & Cioppa 2002). These changes are then reflected by changes in the colour of the conodonts. Well preserved conodonts have a whitish-amber colour and are translucent, whereas conodonts affected by burial and diagenesis become opaque and darker. The colour changes are irreversible and cumulative, and can be correlated with other organic maturity indexes, such as fixed carbon, palynomorph lucency, and vitrinite reflectance (Epstein et al. 1977). These changes reflect the degree of organic metamorphism and thermal maturation, and can provide information about oil and gas occurrences (e.g. Epstein et al. 1977; Symons & Cioppa 2002). Unaltered conodonts (CAI=1) are generally pale yellow with a smooth surface, because they

have had a minimal thermal history and have never been deeply buried (Epstein et al. 1977; Fig. 7).

Figure 7 shows unaltered conodonts from the Kope Formation, Kentucky, USA, studied by Epstein et al. (1977). These conodonts have some colour variation, related to the maturity and the robustness of the individual element, but most of them are pale yellow in colour and have not been heated at temperatures higher than 50 °C. Conodonts affected by increasing temperatures will gradually (more than 50 °C) change colour (Fig. 8). These colour changes have been used for the establishment of the Colour Alteration Index (CAI). The changes are both time and temperature dependent (Epstein et al. 1977; Rejebian et al. 1987). Conodonts having a CAI value of 1.5 are pale brown in colour and reflect the temperature interval 50–90 °C and a fixed carbon range of 55–70%. With increasing temperature the CAI value will increase from 1.5 to 5, reflected by a change in conodont colour from very pale brown (CAI=1.5) through brown to dark brown (CAI=2–3; fixed carbon range *c*. 65–80%) to black colour (CAI=4–5; fixed carbon range *c*. 80–95%). A CAI value of 5 reflects a temperature ≥ 300 °C (Table 2; Epstein et al. 1977; Rejebian et al. 1987; Voldman et al. 2008).
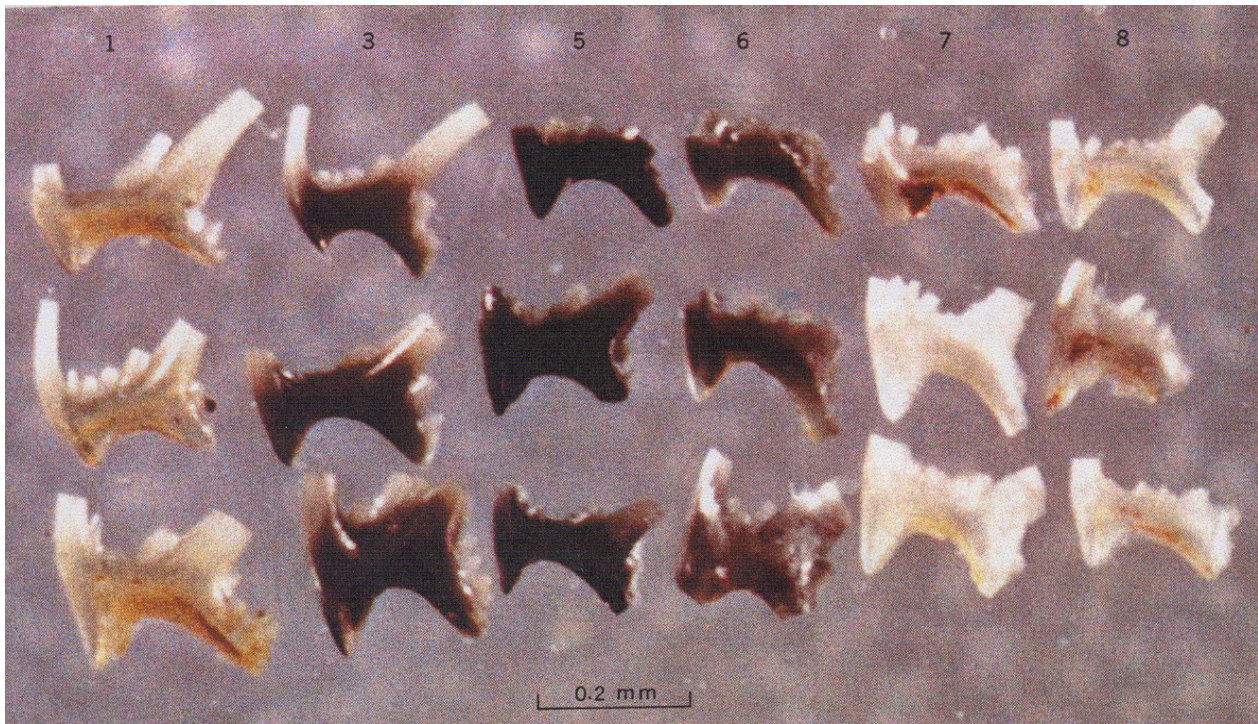
11

Fig. 9. Changes in colour alteration of conodonts. The CAI values are indicated at the top of each column. From Epstein et al. (1977, fig. 6).

At high temperatures or when the CAI have values more than 4.5, conodonts change in surface texture and colour, and their surface changes from smooth and vitreous to pitted and grainy (Epstein et al. 1977; Rejebian et al. 1987). With continued heating, the conodont colour will change from black (chlorite grade) to grey (CAI= 6), then to opaque white (CAI=7), and finally (CAI=8) to crystal clear as a consequence of oxidation of organic matter, release of water and recrystallization (Epstein et al. 1977; Rejebian et al. 1987). A CAI value of 8 corresponds to temperatures over 600 °C (Fig. 9; Epstein et al. 1977; Rejebian et al. 1987; Voldman et al. 2008).

The conodont CAI values can be classified into three zones: the diacaizone, the ancaizone, and the epicaizone. The diacaizone and the ancaizone indicate three maturation stages for organic matter (diagenesis, catagenesis, and metagenesis). The diagenetic stage corresponds to temperatures between 50 to 90 °C and CAI values ranging from 1 to 1.5, reflecting the formation of immature of hydrocarbons (Voldman et al. 2008). CAI values ranging from 2 to 3 indicate the catagenetic stage, during which heavy oil and wet oil are formed (Epstein et al. 1977; Rejebian et al. 1987; Burrett 1992; Marshall et al. 1999; Symons & Cioppa 2002; Voldman et al. 2008).

CAI values ranging from 4.0 to 4.5 indicate the metagenetic stage, which is characterized by dry gas generation. The hydrocarbons become overmature when the CAI value is ≥5 and temperature + 300 °C

Table 2. Correlation of Conodont Alteration Index (CAI) with temperature and hydrocarbon zones.

Conodont colour change	CAI value	temperature	Fixed carbon range	Maturation stage	Hydrocarbon zone
Pale yellow	1	≤50 - 80°C	60%		Immature zone
Very pale brown	1.5	50 - 90°C		Diagenesis	
Brown to dark brown	2	60 - 140°C	55% -70 %		Heavy oil zone- to wet gas zone
Very dark grayish brown to dark reddish brown to black	3	110 - 200°C	70% - 80%	Catagenesis	
Black	4	190 - 300°C	80% - 95%	Metagenesis	dry gas zone
Black	5	+300°C	+95%		Overmature

(Epstein et al. 1977; Rejebian et al. 1987; Burrett 1992; Marshall et al. 1999; Symons & Cioppa 2002; Voldman et al. 2008).

According to Epstein et al. (1977), the colour alteration of conodonts is not solely a function of the temperature but also of the duration of the heating (see appendix I). The relative effect of heating time is, however, less important than that of temperature (Bergström 1980). It has also been suggested that tectonics can influence conodont colour alteration, because high CAI values (and high rank coals) are common in areas with extensive thrusting and folding. Epstein et al. (1977), however, studied Devonian conodonts in the northeastern part of the Appalachian basin, USA, and concluded that tectonics do not affect the colour alteration in conodonts. Intrusives and hydrothermal activity are on the other hand likely to have affected conodonts thermally (e.g. Bergström 1980; Rejebian et al. 1987).

5.1.3 Examples of application

Upper Palaeozoic strata are widely exposed in the Anti-Atlas of Morocco, where they form part of an exposed segment of the Variscan autochthonous deformation zone that is surrounded by flat-lying Cretaceous–Cenozoic deposits (Belka 1991). The Palaeozoic successions in the eastern Anti-Atlas are dominated by clastics with rare volcanic flows (Belka 1991). The Devonian succession largely consists of graptolitic shales and pelagic carbonates (limestones and dolomites). In the Mader and Tafilat basins, it attains a thickness of 2000 and 1000 m, respectively, whereas the pelagic platform deposits are only 100–200 m (Fig. 10; Belka 1991).

The CAI pattern in the Devonian of the eastern Anti-Atlas show values ranging from 3 to 5, indicating burial palaeotemperatures of 120–300 °C. The Mader Basin has the thickest succession, which is reflected by high CAI values (CAI=5). The area situated in the south is covered by Cretaceous deposits and has CAI=4. Other areas along the eastern Anti-Atlas have CAI 2.5–3, indicating a palaeotemperature of 120 °C (Fig. 11; Belka 1991). According to Belka (1991), the CAI pattern is the result of variation in burial depth and loading by a thick Carboniferous succession. The heat-flow regime seems, however, to have been higher than normal, with a geothermal gradient of approximately 50 °C/km (Belka 1991), and his study illustrates the usefulness of conodonts in the study of sedimentary loading and geothermal history of sedimentary basins.

Uppermost Silurian through lowermost Devonian rocks in the Appalachian basin were studied by Epstein et al. (1977). They constructed CAI isograds on the basis of conodonts recovered from limestones. Along the Allegheny front in Pennsylvania, the CAI values vary between 2.5 and 3 (Epstein et al. 1977; Fig. 12). Towards the east, the CAI values generally range between 4 and 4.5 (Epstein et al. 1977), reflect-
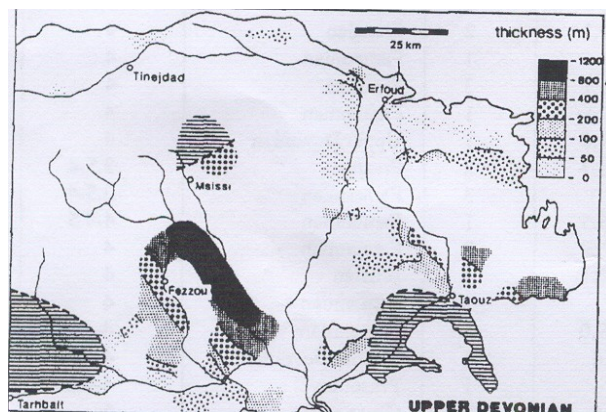

Fig. 10. Thickness of the upper Devonian in the eastern Anti-Atlas in Morocco. From Belka 1991, fig. 3).
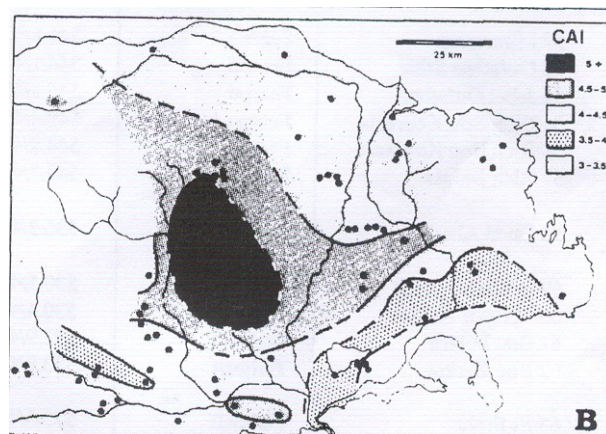

Fig. 11. CAI values in the Lower and Middle Devonian in the eastern Anti-Atlas in Morocco. From Belka (1991, fig. 4).

ing an eastward-thickening of upper Devonian through upper Carboniferous clastic wedges. Oil fields are largely restricted to areas west of the 2 isograd in the surface, whereas gas production is most common in areas where the surface rocks have CAI values as high as 4. Thus, the western part of the southern Appalachian basin has the best potential for oil and wet gas production. To the east there are areas with high potential for dry production, because the CAI values are close to 4.0– 4.5 (Epstein et al. 1977).

The colour alteration index of conodonts has also been used for studies of the maturation and burial history of the Palaeozoic successions in southern Ontario (Legall et al. 1981). CAI values revealed three facies of thermal alteration. The first facies is present in the southwestern part of Ontario and characterized by CAI values from 1.0 to 1.5, reflecting maximum burial temperatures of less than 60 °C (Legall et al. 1981). The second zone has CAI values from 2 to 2.5, reflecting burial temperatures between 60 °C and 90 °C, i.e. the oil and wet gas zone (Legall et al. 1981). The third facies has high CAI values of 3–4 as response to burial temperatures ranging from 90 to 120 °C, i.e. the limit for light oil and wet gas generation,
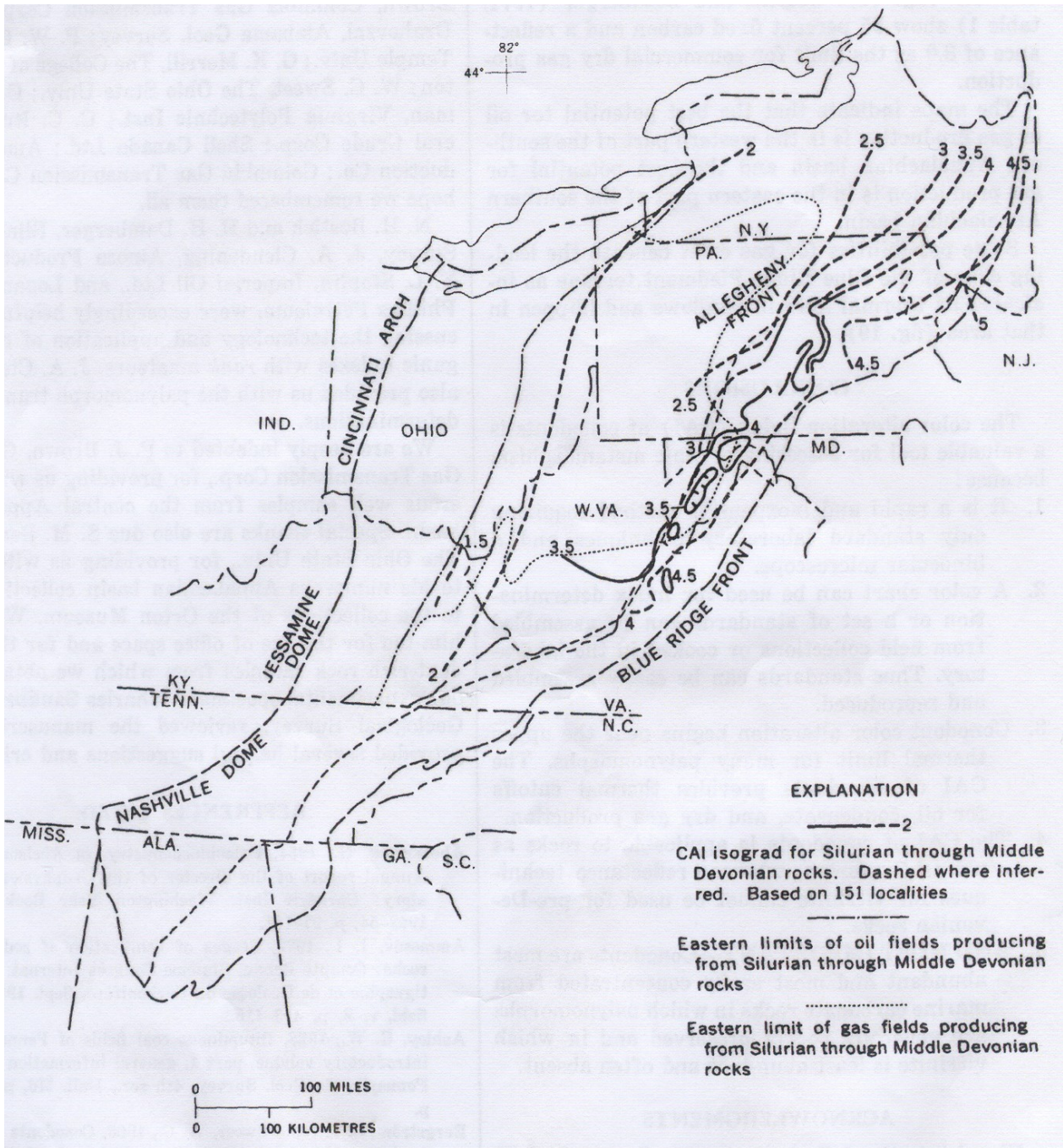
13

Fig. 12. Map of the Appalachian basin showing CAI isograds for Silurian through Middle Devonian strata and the eastern limit of hydrocarbon fields. From Epstein et al. (1977, fig. 20).

and the beginning of dry gas generation. CAI values of 4–5 indicate the dead line for dry gas generation (Legall et al. 1981).

5.2 Graptolites
5.2.1 Introduction

Graptolites are an extinct group of marine, colonial invertebrates that lived from the middle Cambrian to the Carboniferous and built an external skeleton from a variety of proteins, mainly by collagen (Milsom & Rigby 2010). They are generally regarded as a class within the phylum Hemichordata and subdivided into six orders: Dendroidea, Tuboidea, Camaroidea, Crustoidea, Stolonoidea, and Graptoloidea (Clarkson 1998). The orders Dendroidea and Graptoloidea are the most important ones and of prime interest as stratigraphical indicators.

Graptoloids are important for reflectance measurements and maturity studies in Ordovician to lower Devonian strata, whereas dendroids are important for maturity studies of post-Silurian strata (Link et al. 1990; Cardott & Kidwai 1991). Graptolites preserved in shale were generally not affected by bottom currents and scavengers, as opposed to graptolites preserved in most carbonate rocks (Link et al. 1990; Cardott & Kid-

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wai 1991). The graptolite colony is known as the rhabdosome, which consists of thecae (living tubes for the zooid), apertures, a common canal, a virgula, and a periderm (the material composing the rhabdosome) (Link et al. 1990; Clarkson 1998).

Graptolites have distinct morphological features and graptolite fragments show anisotropy, lamellar structure and the surface may be granular or nongranular (Link et al. 1990). Granular fragments are found in carbonate rocks and composed of the exoskeleton or common canal, whereas nongranular fragments are common in shales and are hard and brittle (Link et al. 1990; Goodarzi et al. 1992). Most graptolites were planktonic and settled on the sea floor with their largest dimensions horizontal (e.g. Goodarzi & Norford 1989; Fig. 13). Thus, maximum surface for determination of their optical properties is commonly in sections parallel to bedding (Goodarzi & Norford 1989). Graptolites are used in petroleum exploration because they are important thermal indicators and can be used for determining the degree of maturity of the host rock (e.g. Goodarzi 1985; Goodarzi & Norford 1989; Link et al. 1990). Their reflectance ($R_\circ$) can also be correlated with vitrinite reflectance and the colour alteration index in conodonts (CAI).

5.2.2. Graptolite reflectance

The reflectance of graptolites is an important thermal indicator for lower Palaeozoic rocks in which vitrinite and spores are absent. The maximum reflectance increases gradually with depth of burial, and hence the reflectance can resolve the burial history and the maturity Palaeozoic successions. Graptolites were first used as thermal indicators by Kurylowicz et al. (1976), who noted that the reflectance and optical properties of graptolite fragments in Ordovician strata of the Amadeus basin in Australia closely resemble those of the vitrinite maceral group of coal (see, e.g., Cardott & Kidwai 1991). Since then, graptolites have been widely used for thermal maturation studies in a wide range of areas. Subsequent studies have, however, shown that the graptolite maturation curve differs from the maturation curve of vitrinite (Cardott & Kidwai 1991).

The reflectance is determined by measuring the percentage of incident white light reflected from a polished surface of the periderm (Cardott & Kidwai 1991). It is commonly evaluated by measuring the minimum, maximum and random reflectance of several specimens (Link et al. 1990). Random reflectance is, however, not a reliable method because graptolites are biaxial (Link et al. 1990). The accuracy of the reflectance measurements is dependent on the sample orientation, for instance the orientation of the section with respect to the bedding, and the morphology and structure of the graptolite (e.g. Goodarzi & Norford 1989; Link et al.1990; Goodarzi et al. 1992). Gorter (1984) noted that the reflectance of graptolites and vitrinite are more or less equivalent when the CAI is 1.5–2.0. At CAI ≥3 graptolite reflectance is much

higher than vitrinite reflectance (e.g. Gorter 1984; Cardott & Kidwai 1991).

Two types of graptolite fragments are used to estimate the light reflectance, granular and nongranular. The granular type is found in carbonates and characterized by a lower reflectance than the nongranular types and has a lath-like morphology with a mosaic of

Fig. 13. Early Silurian graptolites from Litohlavy in the Skryje-Týřovice Basin, Barrandian area, Czech Republic. The silvery appearance of the rhabdosomes indicates substantial heating and coalification of the periderm, either due to considerable burial depth or proximity to intrusives. Photo: Per Ahlberg, June 2010.

very fine grains (Link et al. 1990). Goodarzi & Norford (1985) and Link et al. (1990) suggested that some granular fragments could be part of a common canal. Nongranular fragments are found in shales. They are characterized by a higher reflectance and appear as thin blades to the sections normal to bedding, but show a blocky morphology on sections cut parallel to the bedding planes (Link et al. 1990). At low maturation levels, the difference in reflectance between these two types is very small (Goodarzi & Norford 1985). Studies from the Palaeozoic of Artic Canada, have for instance shown that the difference between granular and non-granular fragments is 0.07% $R_\circ$ (Goodarzi et al. 1992). At higher levels of maturation, only nongranular graptolite fragments are present (Goodarzi et al. 1992).

Maximum reflectance ($R_{\circ\ max}$) is the most reliable method for evaluating the thermal maturity of
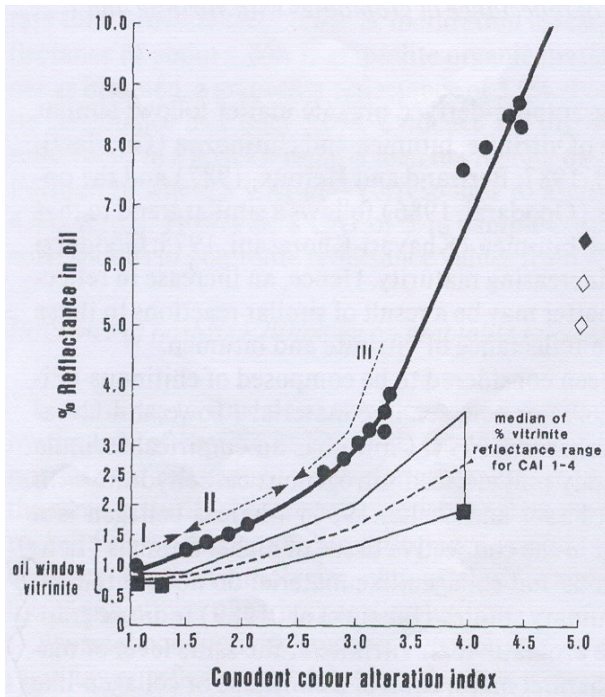
Fig. 14. Vitrinite and graptolite reflectance data plotted versus conodont colour alteration index. From Link et al. (1990, fig. 8).

graptolites and should be measured parallel to the bedding (Link et al. 1990; Goodarzi et al. 1992). Minimum reflectance ($R_{\circ\ min}$) is measured normal to bedding (Link et al. 1990). The thermal maturity of sedimentary successions and the potential for oil generation can be estimated by a correlation of graptolite reflectances and conodont alteration indices (e.g. Goodarzi & Norford 1985, 1989). Silurian graptolites from the Chelm borehole in Poland have a reflectance (%R_{oil} 1.1–1.65) corresponding to CAI values of 1.5–2, indicating the catagenetic phase in the generation of petroleum, whereas graptolites from the Lower Ordovician Flagabro drill core, southernmost Sweden, have %R_{oil} 3.20–3.30, which correspond to CAI 3 and indicate the metagenetic stage and overmature organic matter (Goodarzi & Norford 1989). Graptolites from overmature rocks in the Ordovician of the Akutlak creek in western Canada, exhibit higher reflectance (%R_{oil} 7.8–8.9) corresponding to CAI 4–5 (Goodarzi & Norford 1989). Figure 14 shows the correlation between graptolite reflectance and CAI values. The oil window has $R_{\circ\ max}$ 1.2–2.2 and CAI 1.5–2.5. $R_{\circ\ max} \geq 3$ indicates an overmature stage and the dry gas window (Bertrand 1990; Fig. 14). Goodarzi & Norford (1985) noted that graptolite reflectance is more useful to estimate the maturation than the conodont alteration index because it can be used at high levels of maturation (e.g. Goodarzi & Norford 1989).

The lithology of the host rock can influence the reflectance of graptolites. Graptolites in limestone have, for instance, lower values than graptolites from shales. This is because limestones are generally more permeable and affected by weathering, oxidation and/or dissolution (Link et al. 1990).

5.3 Scolecodonts
5.3.1 Introduction

Scolecodonts represent jaws of polychaete annelid worms. They range in age from latest Cambrian to Recent and are common microfossils in Palaeozoic rocks (e.g. Eriksson 2000). Scolecodonts can be found in both carbonates and shales, and commonly occur as isolated tooth-like structures. They can, however, also be found as fragments or occasionally as jaw apparatuses. They are most common in Ordovician–Devonian shallow marine beds and can be extracted from the host rock by dissolution in acid (e.g. Goodarzi & Higgins 1987; Armstrong & Brasier 2005).

Scolecodonts are resistant and preserved rather easily because they consist of organic matter, including hard proteins, along with a variety of inorganic compounds (Goodarzi & Higgins 1987; Bertrand 1990; Eriksson 2000). Their chemical composition is similar to that of graptolites and chitinozoans. In general, scolecodonts have a granular to non-granular surface and their colour range from dark grey to light grey (Goodarzi & Higgins 1987). Scolecodonts are primarily used in biostratigraphy and palaeoecology, but have also been used in thermal maturity studies because their reflectance increases with increasing temperature (Goodarzi & Higgins 1987; Bertrand & Héroux 1987; Bertrand 1990).

5.3.2 Scolecodont reflectance

The optical properties of scolecodonts show similar trends to those of graptolites and vitrinite, and can be used to determine the degree of maturation. With increasing burial depth, the chemical composition and the reflectance of scolecodonts will change (Goodarzi & Higgins 1987). The colour of scolecodonts also varies with temperature, from yellowish through brown, and finally black (e.g. Goodarzi & Higgins 1987). The degree of maturation of organic matter is determined by measuring the reflectance in air and oil and the refractive and absorptive indices (e.g. Goodarzi & Higgins 1987). These optical properties increase with the Conodont Alteration Index, which indicate that scolecodonts are useful for maturation studies (Goodarzi & Higgins 1987; Bertrand 1990).).

Scolecodont reflectance increases with increasing CAI values and corresponds to vitrinite reflectance. Scolecodont reflectance has similar values to those of graptolite in carbonate matrix, whereas graptolite reflectance in shale matrix shows higher values than the scolecodont reflectance (e.g. Goodarzi & Norford 1985; Goodarzi & Higgins 1987). Scolecodonts with a reflectance between 0.35 and 0.84% corresponds to a CAI value of 1.5–2. 5 and a vitrinite reflectance of 0.8–1.1, indicating that the sediments are within the oil window (e.g. Goodarzi & Higgins 1987; Fig, 15). Scolecodonts having a high reflectance equal
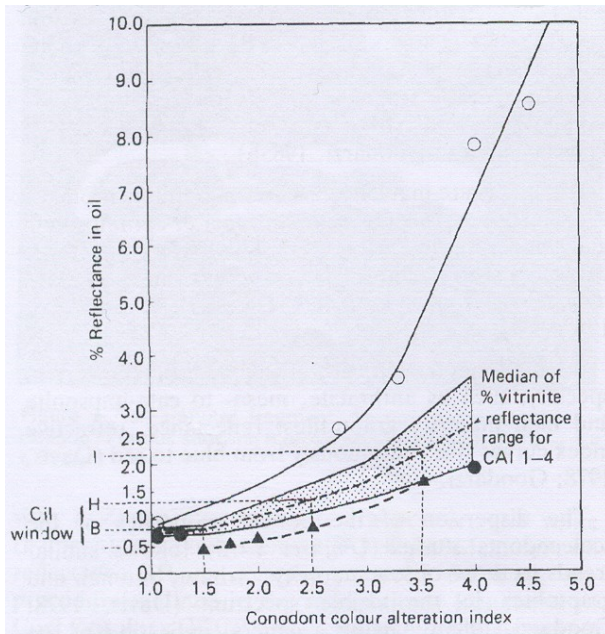
16

Fig. 15. Reflectance of scolecodonts in oil versus conodont alteration index. From Goodarzi & Higgins (1987, fig. 13).

to 1.63% can be correlated to CAI values ≥3 and a vitrinite reflectance equal to >1.65%. These high values indicate the dry gas window in which the organic matter is overmature (e.g. Goodarzi & Higgins 1987).

5.4 Spores and pollen
5.4.1 Introduction

Spores and pollen are produced by land plants and are generally very common in sedimentary rocks. Spores and pollen are studied in a dispersed state and form the basis of palynology as originally proposed by Hyde & Williams (1944; cf. Taylor 1996). The taxonomical use of spores and pollen follows the rules of the International Code of Botanical Nomenclature (ICBN).

Spores are produced in the life cycle of the "lower plants", such as bryophytes and vascular cryptogams ("pteridophytes"), and have been recorded from Mid Ordovician to the present (Rubinstein et al. 2010). Bryophytes are structurally intermediate between green algae and vascular plants (the Tracheophyta) and are very moisture dependent (e.g. Playford & Dettmann 1996). They have simple leaves and root-like structures, and the sporophyte generation is inconspicuous compared to the gametophyte generation (Playford & Dettmann 1996; Armstrong & Brasier 2005). The Pteriodophyta has no classificatory significance and includes free-sporing vascular plants such as ferns and fern allies. This complex group of vascular plants are important in paleopalynology (e.g. Playford & Dettmann 1996).

Pollen is produced by "higher plants" (angiosperms and gymnosperms) and first appears in the upper Devonian. The development of pollen grains differs in detail between gymnosperms and angiosperms (Jarzen & Nichols 1996). Angiosperm pollen

appeared in the early Cretaceous and has high diversity. Both spores and pollen are frequently used in correlation and biostratigraphy, but they are also valuable for palaeoenvironmental, palaeoecological, and palaeobiogeographical studies (e.g. Armstrong & Brasier 2005). Spores and pollen are used to decipher the flora within a spatial framework (Taylor 1996; Jarzen & Nichols 1996; Playford & Dettmann 1996). They are also widely utilized in hydrocarbon exploration through thermal maturity studies.

The taxonomy of spores and pollen is based on the morphology such as their shape, size, apertures and wall structure. The shape of spores depends on the meiotic divisions of the spores from the mother cell (Armstrong & Brasier 2005). When the mother cell divides, it leads to a tetrad comprising four smaller cells. In this case the spores have tetrahedral tetrads and each spore is in contact with all three of its neighbours on the proximal face (Armstrong & Brasier 2005). When the mother cell splits into two cells, the tetrad will be of a tetragonal shape and may resemble the segments of an orange (Traverse 1988).

The morphology of gymnosperm pollen differs in several respects from angiosperm pollen. Gymnosperm pollen varies from small, spherical and inaperaturate to large, bisaccate and ornamented (e.g. Jarzen & Nichols 1996). Some pollen grains are spherical to subspherical, and have a smooth or granular surface with a distal furrows or sulcus (e.g. Jarzen & Nichols 1996). Several gymnosperm genera produce saccate pollen bearing one, two or three air sacs attached to a central body (e.g. Jarzen & Nichols 1996). Angiosperm pollen shows a greater variation than gymnosperm pollen. The angiosperm pollen grains may be aperturate or inaperaturate and provided with one or more pores (monoporate, diporate, triporate etc.; e.g. Jarzen & Nichols 1996).

The colour is related to the degree of maturation and spores and pollen grains can be used as paleothermometres. The colour changes as a response to heating and depth of burial. This colour change has been recognized since 1920 (Ujiié 2001). The spore colour is a function of the decomposition of organic matter due to coalification and carbonization (e.g. Gray & Boucot 1975).

5.4.2 Spore and pollen colour index

The spore and pollen colour index is along with vitrinite reflectance the most common method used for estimate the degree of organic maturity. Increased temperatures result in rapid chemical and physical alteration of spores and pollen, and the carbon content increases relative to loss of oxygen and hydrogen (Marshall 1991). This chemical and physical change will lead to a change in the colour of spores and pollen (Marshall 1991). This colour change is called the Thermal Alteration Index (TAI), which is influenced by temperature, burial depth and time.
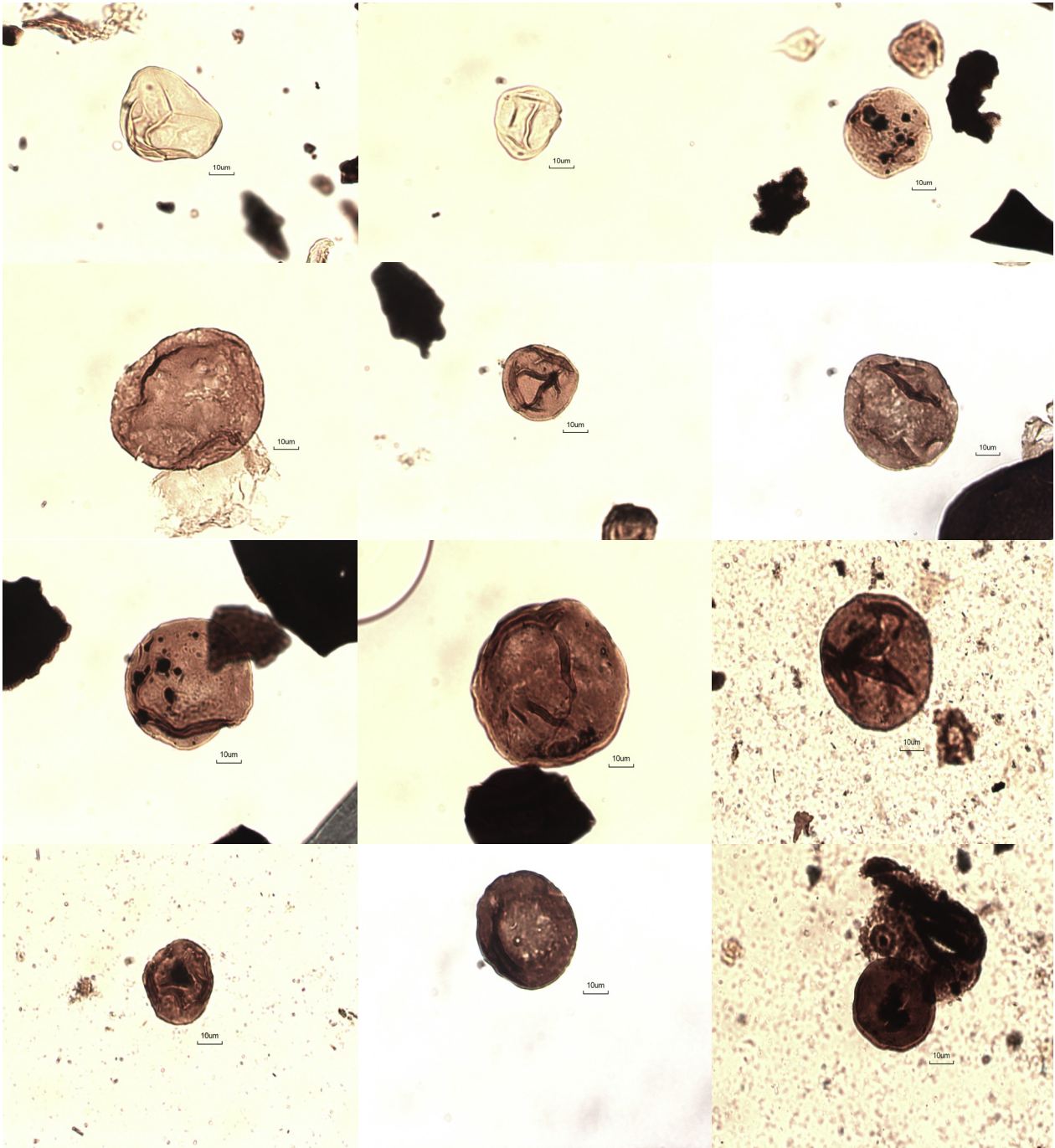
Two microscopic methods are used for measur-

Fig. 16. Late Silurian spores from Scandinavia showing a range of colours.
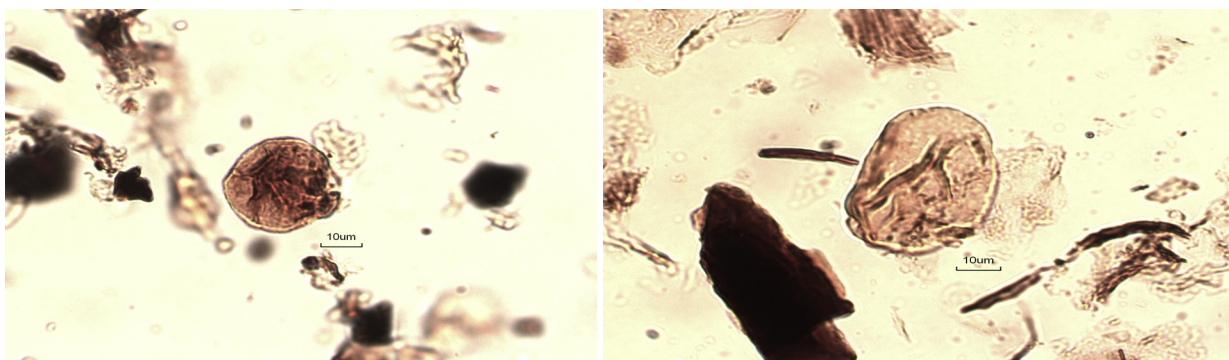
Fig. 17. Late Cretaceous spores/pollen from New Zealand showing a range of colours.

ing the TAI of spores and pollen. One method is simply to distinguish their morphology and colour by using the eyes, and the other method is to establish the translucency of organic matter by using photoelectric measurements (Ujiié 2001). Assessment of the colour

Table 3. Colour changes within spores and the Spore Colour Index (SCI). Modified from Marshall (1990, fig. 2).

Spores colour	SCI	
Pale yellow	1	
Pale yellow- lemon yellow	2	
Lemon yellow	3	
Golden yellow	4	
Yellow orange	5	
Orange	6	
Orange brown	7	
Dark brown	8	
Dark brown- black	9	
black	10	

change is generally difficult because the colour can vary widely within a single assemblage according to several factors, such as variations in composition and wall thickness, degree of oxidation and degradation, and occasionally the presence of the reworked specimens (El Beialy et al. 2010). The Spore Colour Index

(SCI) ranges from 1 to 10 and reflects a colour gradation from colourless or pale yellow to black (Marshall 1990; Utting & Hamblin 1991; Table 3). The TAI is assessed from the colour of spores and pollen before oxidation treatment of the samples, and by using a five point scale (Marshall 1990; Utting & Hamblin 1991; Table 4). A transparent and pale yellow colour indicates TAI 1, medium orange TAI 2, dark brown TAI 3, black TAI 4, and vitreous black and brittle spores and pollen indicate TAI 5 (Utting & Hamblin 1991; Figs.. 16, 17).

TAI 1 corresponds to the early stage of diagenesis of organic matter and reflects only minor chemical changes. It indicates temperatures less than 50 °C and the immature zone of hydrocarbons (Marshall 1990; Utting & Hamblin 1991; El Beialy et al. 2010). Thus, fossil specimens of long-ranging taxa, such as *Botryococcus*, *Triplansoporites*, *Biretisporites*, and *Cocavisporites*, with a lustrous pale yellow colour indicate a thermally immature zone (El Beialy et al. 2010). As a response to increasing temperatures spores and pollen will change from yellow to medium orange (TAI 2), indicating the later stage of diagenesis and the beginning of the oil window. TAI 2 is roughly equivalent to vitrinite reflectance $\%R_{\circ} = 0.5$ and CAI 1.5, suggesting temperatures more than 60 °C (Marshall 1990; Utting & Hamblin 1991). TAI 2+ indicates the peak of wet gas generation. With increasing temperatures the TAI increase to 3 and the spores and pollen have a dark brown colour, indicating the oil "death" line and the peak of the dry gas generation in terms of petroleum zones (Utting & Hamblin 1991). TAI 3 is approximately equivalent to vitrinite reflectance $\%R_{\circ} =$

Table 4. Correlation between colour changes in spores and pollen, Thermal Alteration Index (TAI) and hydrocarbon generation zones. Based on Marshall (1990, fig. 2) and (Uttin & Hamblin 1991, table 1).

Spores and pollen colour	Thermal alteration index	Petroleum generation
Pale yellow- yellow	1	Immature kerogen
Yellow to light orange –medium orange	2	Oil window
Dark brown	3	Gas window
Brownish black to black colour	4	wet gas limit – dry gas preservation
Vitreous black- fossils brittle	5	Dry gas preservation limit

1.20 and CAI = 2, suggesting the catagenetic stage and temperatures more than 120 °C.

When the temperature increases to be more than180 °C, the spore and pollen colour will be brownish black to black, indicating TAI 4 and the wet gas floor (Utting & Hamblin 1991). TAI 4 corresponds to a vitrinite reflectance equivalent to 2.0–3.0 and CAI = 4–5 (Marshall 1990; Utting & Hamblin 1991; Voldman et al. 2008). Vitreous black to brittle spores (TAI 5) indicate the dry gas preservation limit and the over-mature zone of kerogen (metagenesis; Marshall 1990; Utting & Hamblin 1991; Voldman et al. 2008).

Spore colouration has been applied for thermal maturity studies in a wide range of areas around the world, particularly in areas with near-shore marine, fluviatile and lacustrine successions. Utting & Hamblin (1991) showed, for instance, that spore colouration could be used for studying the hydrocarbon potential of the Lower Carboniferous Horton Group of Nova Scotia, Canada. The thickness of this group varies considerably and has TAI values ranging from 1 to 5, and they (Utting & Hamblin 1991) were able to distinguish the oil and gas windows within a succession of fluviatile and lacustrine siliciclastics.

5.5 Acritarchs
5.5.1 Introduction

Acritarchs are organic-walled, eukaryotic microfossils of unknown and probably polyphyletic affinities (Armstrong & Brasier 2005). Most of them are considered to represent vegetative cells and resting or reproductive cysts of unicellular green algae or dinoflagellates, cyanobacterial envelopes and perhaps fungi (Strother 1996; Talyzina & Moczydłowska 2000; Vecoli & Hérrissé 2004; Butterfield 2005). They range in age from the upper Palaeoproterozoic to the Recent, and reached their acme during the Palaeozoic (e.g. Strother 1996; Armstrong & Brasier 2005).

Acritarchs have a diverse morphology that sometimes can reflect their origin (Strother 1996). They generally consist of a hollow vesicle that can be ornamented and provided with processes (e.g. Strother 1996; Armstrong & Brasier 2005). The shape of the vesicle varies within wide limits and the exterior surface can be smooth, granular, reticulate or spinose (Strother 1996; Armstrong & Brasier 2005). The classification is informal and based on morphotypes (Willman & Moczydłowska 2007). Three morphological groups can be recognized: (1) acritarchs lacking processes or crests, (2) acritarchs with crests but lacking processes, and (3) acritarchs with processes, with or without crests (Armstrong & Brasier 2005).

Acritarchs are most common in marine shales and mudstones, and can be used for palaeoecological interpretations (e.g. Vecoli & Hérrissé 2004; Armstrong & Brasier 2005). Lagoonal, nearshore, and deep offshore facies are generally characterized by low diversity acritarch assemblage facies (Armstrong & Brasier 2005). Acritarchs are widely used for biostratigraphy and correlation of upper Precambrian and Palaeozoic strata (e.g. Strother 1996; Armstrong & Brasier 2005).

Acritarchs can be used in petroleum exploration by determining the degree of maturation of organic matter, because their colour changes as a response to increasing temperature and burial depth. Thus, the Acritarch Alteration Index (AAI) can provide important information on the thermal history of sedimentary basins.

5.5.2 Acritarch Alteration Index

Acritarch Alteration Index (AAI) is a good indicator for estimating the maturation at low temperatures within the oil window (Legall et al. 1981). With increasing temperature and depth of burial, acritarchs undergo similar changes in wall colour as pollen and spores (Marshall 1991). They are important in successions where spores and vitrinite are absent, for instance in Lower Palaeozoic successions.

Acritarchs change progressively in colour from light yellow and translucent, through pale yellow, yellow orange, orange, brown, to black. The AAI consists of a five point scale based on the colour changes (Legall et al. 1981; Marshall 1991). Translucent to light yellow (AAI = 1) and light yellow to pale yellow (AAI = 2) indicate the immature hydrocarbon zone (temperatures less than 70 ℃). Pale yellow to orange acritarchs (AAI = 3) indicate the upper part of the oil and wet gas window, whereas orange to dark brown acritarchs (AAI = 4) indicate the peak of the oil and wet gas window. Black acritarchs (AAI = 5) are indicative of the lower oil and wet gas window and the upper part of the dry gas window (Table 5; Legall et al. 1981; Marshall 1991). Figure 18 shows different species of acritarchs from Sweden, such as *Baltisphaeridium* sp., *Ammonidium* sp. and *Diexallophasis* sp., with different colours and values of the AAI.

The Acritarch Alteration Index (AAI) can be correlated to the Conodont Alteration Index (CAI). Acritarchs can, however, only be used for estimating maturation at low temperatures, whereas conodonts can be used for estimating maturation that range from low to high temperatures. AAI values lower than 2 corresponds to CAI values less than 1.5, and AAI = 3–4 approximately corresponds to CAI = 2.5, indicating the oil window and the first stage of wet gas generation (Marshall 1991, fig. 4; Williams et al. 1998, fig. 4). AAI = 5 corresponds to CAI = 2.5–4, indicating the wet gas limit and the dry gas zone (Marshall 1991, fig. 4; Williams et al. 1998, fig. 4). At higher temperatures acritarchs cannot be used for maturation studies, whereas conodonts are important for maturation studies within the dry gas window.

Legall et al. (1981) and Legall & Barnes (1980) studied the thermal history of the Palaeozoic of southern Ontario by using the colour alteration indices of acritarchs and conodonts. Based on CAI values, three thermal alteration zones of organic maturation were
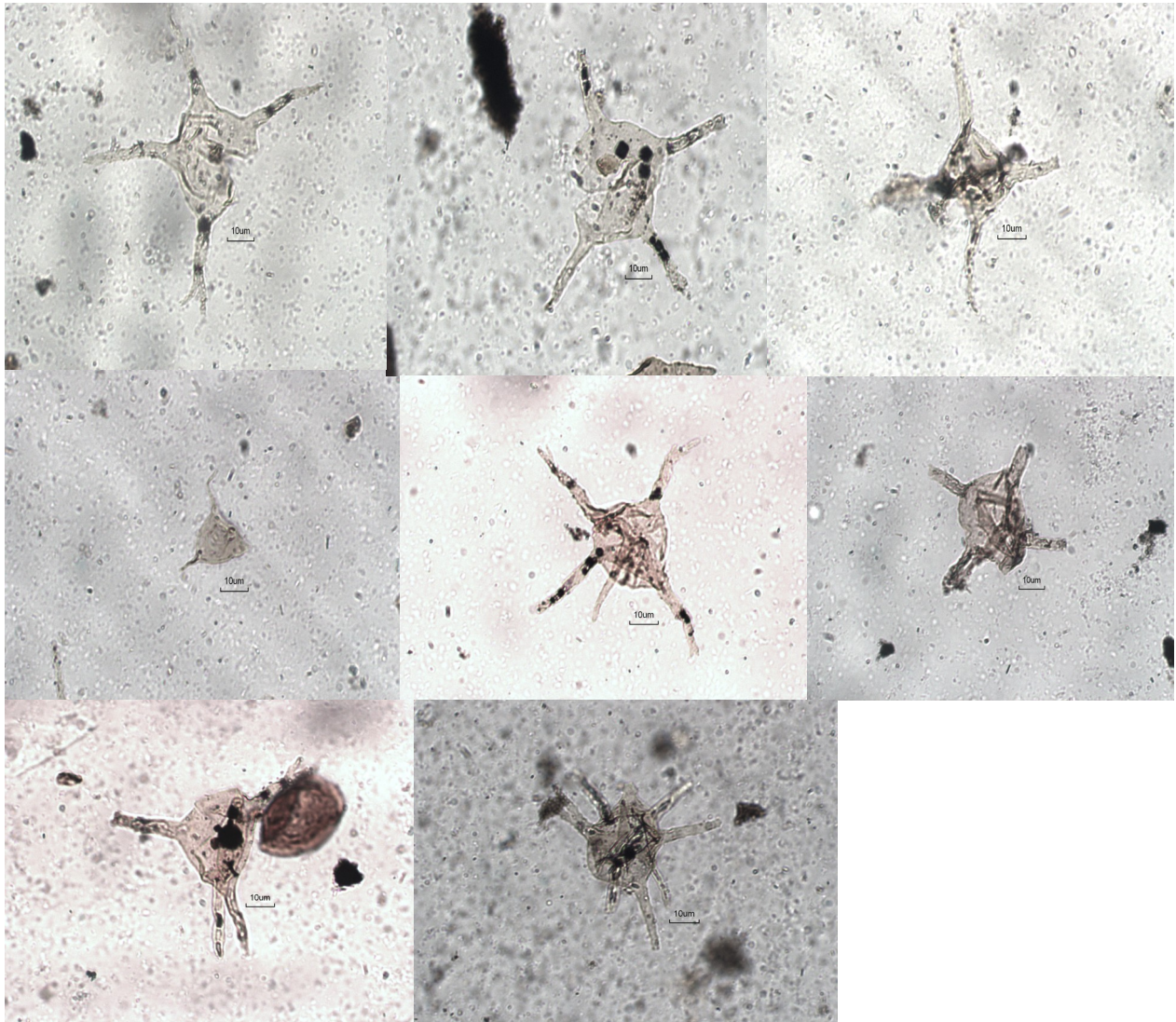
Fig. 18. Late Silurian acritarchs from Scandinavia showing a range of colours.

Table 5. Correlation between colour changes in acritarchs, Acritarch Alteration Index (AAI) and hydrocarbon generation zones. Based on Marshall (1990, fig. 4).

Acritarchs colour change	Acritarch alteration index (AAI value)	Temperature	Petroleum zone
Colourless to light yellow	1	≤60˚C	Immature kerogen
Light yellow to pale yellow	2		Heavy to light oil
Pale yellow to orange	3	90 ˚C	
Orange to dark brown	4		And first of wet gas zone
Black colour	5	120˚C -185˚C	Wet gas limit and first of dry gas zone

recognized. The boundaries between these three zones were refined by the use of acritarchs. The first zone extends from the top of Palaeozoic sedimentary succession and has the AAI values changes gradually from 2 to 4 in the direction of depth of burial (Legall et al. (1981). The second zone shows degraded acritarchs with AAI values equal to 5 (Legall et al. 1981; Legall & Barnes 1980).

5.6 Vitrinite

5.6.1 Introduction

Vitrinite is a primary component of coal and an important maceral group. It is derived from lignin, cellulose, and tannins of vascular plants, especially from the bark and wood tissues, and can also occur as dispersed organic particles in various sedimentary rocks (Dow 1977; Mukhopadhyay 1994). It can also be considered humic or structured organic matter (Dow 1977). Vitrinite has diverse morphological characteristics and can be subdivided into subgroups and macerals, depending on the morphology, relief on polished surfaces and optical properties. It is formed as result of chemical and physical alteration of biopolymers of plants when temperature and pressure increase through time, a process known as maturation or coalification (Mukhopadhyay 1994).

The vitrinite group can be subdivided into three subgroups based on the morphology, diagenetic characteristics and texture: telinite, collinite and detrinite (Mukhopadhyay 1994). Telinite macerals represent fragments or the cell-wall material of land plants, whereas the collinite macerals have an amorphous appearance and represent the organic substance that filled the cell cavities (Mukhopadhyay 1994; Hunt 1995). The detrinite macerals results from smaller fragments, which are occasionally partially gelified (Dow 1977; Mukhopadhyay 1994). Collinite includes four maceral types: corpocollinite, telocollinite, gelocollinite and desmocollinite (Mukhopadhyay 1994). Desmocollinite has a lower reflectance than other collinite macerals.

Hunt (1995) distinguished between two types of vitrinite in coals, organic-rich shales and siltstones: vitrinite-1, which has a low hydrogen content and includes telinite and telocollinite; vitrinite-2, which has a high hydrogen content and includes desmocollinite, heterocollinite and degradinite. For maturity measurements, high-grey vitrinite-1 macerals are used rather than the low-grey vitrinite-2 macerals (Hunt 1995).

5.6.2 Vitrinite reflectance

Vitrinite reflectance is used to determine the maturation of organic matter in sedimentary rocks (e.g. Fermont 1988; Ismail & Shamsuddin 1991; Mukhopadhyay 1994; Hunt 1995). The maturation processes include chemical transformations, such as decarboxylation and dehydroxylation, and physical manifestations, of which the most important one is the increase in reflectance (Mukhopadhyay 1994; Hunt

1995; Huang 1996). Teichmüller (1958) used vitrinite reflectance in order to measure the coal rank, and she was the first one to use dispersed vitrinite particles in shales to extend the rank parameters to successions that did not possess coal deposits.

Two techniques are used for preparing samples to measure the reflectance of vitrinite (Hunt 1995). The first one removes the mineral matter with hydrofluoric and hydrochloric acids. The residue is mounted in epoxy and then polished. The second technique is based on whole-rock polished pellet mounts or thin sections, rather than the kerogen concentrate (Hunt 1995, p. 366). For both techniques, a reflecting-light microscope is used to measure the degree of reflectivity.

Vitrinite reflectance is one of the most important methods in petroleum exploration, because it extends over a longer maturity range than other methods, and can generally be used for measuring high palaeotemperatures of petroleum source rocks (e.g. Carr & Williamson 1990; Ismail & Shamsuddin 1991; Mukhopadhyay 1994; Hunt 1995). Vitrinite is most abundant in shales and siltstones, and restricted to deposits ranging in age from the Devonian to the Quaternary (Epstein et al. 1977).

Vitrinite has several advantages for measuring thermal maturity: (a) it is present in almost all organic-rich sedimentary or metasedimentary rocks; (b) it appears homogenous under the incident light microscope; (c) it shows uniform changes with increasing of temperature and pressure; (d) a large number of analyses can be made in a relatively short time; (Mukhopadhyay 1994; Hunt 1995).

Vitrinite reflectance is expressed as $\%R_o$, which refers to the percentage of incident light that is reflected back through the microscope (e.g. Mukhopadhyay 1994; Hunt 1995). The maximum $\%R_o$ value is 15 (for graphite). Reflectivity largely increases with temperature and time. The measurements and the interpretation of the values require a skilled operator or an experienced petrologist, because the samples may contain caved and recycled material. It may also vary from the presence of pyrite or asphalt (Selley 1985; Hunt 1995). R_o values lower than 0.5 indicate temperatures lower than 50 °C. R_o values ranging from 0.5 to 1.5 indicate the oil window, and R_o values 1.6–2.6 indicate the gas window (Selley 1985; Mukhopadhyay 1994; Hunt 1995; Voldman et al. 2008). R_o values 2.0–3.0 suggest the dry gas hydrocarbon zone (Selley 1985; Mukhopadhyay 1994; Hunt 1995; Voldman et al. 2008). Vitrinite with values higher than around 3.0 indicate overmaturation and a zone devoid of hydrocarbons (Selley 1985; Mukhopadhyay 1994; Hunt 1995; Voldman et al. 2008).

The principal maceral of vitrinite-1 is telocollinite, and that of vitrinite-2 is desmocollinite (Hunt 1995). Telocollinite has a higher reflectance than desmocollinite. Thus, at the same rank, the difference between them can be 0.1–0.3$\%R_o$, with the wider differences at increasing burial depth (Buiskool Toxo-

peus 1983).

The reflectance increases with maturation because the molecular structure of vitrinite changes with temperature. This is because vitrinite consists of clusters of condensed aromatic rings, and with increasing temperature these clusters combine into larger, condensed rings (Hunt 1995, p. 365). The chemical reactions and the maturation changes associated with the increase in temperature are irreversible (Hunt 1995, p. 366).

The hydrogen content causes two problems in vitrinite reflectance. The first one is limiting the reflectance readings to the low-hydrogen vitrinite-1 group and eliminating readings of caving, mud additives, recycled vitrinite and high-hydrogen vitrinite-2 group macerals (Hunt 1995, p. 507). The use of a reflectogram showing the reflectance of all macerals in the sample can minimize this problem (Hunt 1995). The second problem is due to the chemical difference between vitrinite in humic and sapropelic kerogen. Thus, vitrinite in marine shales or lacustrine coal generally shows a lower reflectance than the low-hydrogen vitrinite in humic coals (Mukhopadhyay 1994; Hunt 1995). The lower reflectance of high-hydrogen vitrinite in shales is known as *suppressed vitrinite* (Fermont 1988; Mukhopadhyay 1994; Hunt 1995). The suppressed vitrinite values cannot be easily corrected. Mukhopadhyay (1994) discussed, however, the correction of suppressed vitrinite reflectance to the level of standard reflectance in, for instance, telocollinite grains. He also discussed the two most significant methods for calibration of the reflectance trends.

The lower reflectance of suppressed vitrinite can also be caused by factors, such as: (1) lithological variation and differences in thermal conductivity and heat capacity, (2) abundance of liptinitic maceral in vitrinite-poor source rocks, (3) variation in bacterial degradation of kerogen in the sediment, and (4) impregnation with migrated oil or generated bitumen (Mukhopadhyay 1994).

6 Concluding remarks

Fossils are invaluable for assessing the maturity of kerogen and determining whether or not potential source rocks have matured sufficiently to generate petroleum. They can also be used to determine whether source rocks are oil or gas prone, or whether they are overmatured and barren (Selley 1985). The most important biological indicators for maturation studies are conodonts, graptolites, scolecodonts, spores and pollen, acritarchs, and vitrinite (Fig. 19). The oil window is indicated by CAI = 1.5–2, TAI = 2–3, R_o (vitrinite) = 0.5–1.5, $\%R_{max}$ (graptolite) = 1.2–2.2, AAA = 3–4 and scolecodont reflectance 0.35–0.84 (Fig. 19). The gas windows can be detected by CAI = 3–4.5, TAI = 3–5, AAA = 5, R_o (vitrinite) = 1.5–3, $\%R_{max}$ (graptolite) = 2.2–3 and a scolecodont reflectance of 0.83–1.63.

Pollen and spore colouration and vitrinite re-

flectance are generally the most reliable indices. Palynomorphs, such as acritarchs, spores and pollen, are most common in clastic rocks and provide accurate assessment of low temperatures (the thermal interval below 160 °C) and the hydrocarbon generation zones, whereas conodonts are common in marine carbonates and provide a better assessment of late-stage maturation of organic matter (e.g. Epstein et al. 1977).

Spores and pollen can be used in strata ranging in age from the Silurian to Recent, and can be used to assess the thermal alteration at low temperatures, even below 50 °C, whereas vitrinite is most abundant in argillaceous clastic rocks of Devonian through Quaternary age (e.g. Epstein et al. 1977). Vitrinite can be used to estimate the degree of maturation of organic matter at high temperatures, even to the graphite-grade metamorphic facies (e.g. Epstein et al. 1977; Hunt 1995). Conodonts are used as thermal indicators of marine strata ranging in age from the Cambrian to the uppermost Triassic (Epstein et al. 1977; e.g. Armstrong & Brasier 2005). They can provide information about thermal alteration at high temperatures, extending to the garnet-grade metamorphic facies corresponding to temperatures as high as 500 °C, a temperature far above the destruction of organic-walled palynomorphs (Epstein et al. 1977; Rejebian et al. 1987; Voldman et al. 2008).

Because of their abundance and diversity, spores and pollen are good indicators in lacustrine and near-shore marine Mesozoic and Cenozoic successions, whereas conodonts and graptolites are confined to marine Palaeozoic strata. Thus, conodonts are common in sedimentary rocks in which pollen and vitrinite are virtually absent.

Graptolites are most common in fine-grained clastics of Ordovician and Silurian age. Their maturation follows that of vitrinite, although a direct correlation is not possible because vitrinite is absent in Lower Palaeozoic offshore strata (Goodarzi & Norford 1989). Graptolite reflectance at higher levels of maturation appears, however, to be greater than that of vitrinite reflectance, whereas the reflectance of graptolites at low levels of maturation seems to be lower than that of vitrinite. At high temperatures, graptolites may be more sensitive thermal indicators than conodonts, whereas the correlation between the reflectivity of graptolites and CAI is excellent at lower maturation levels (Goodarzi & Norford 1989). Link et al. (1990) noted that graptolites in limestones has a lower reflectance than graptolites in shales from the same stratigraphical interval. This is probably a result of carbonate dissolution and oxidation of the limestones (Link et al. 1990).

Scolecodonts are common in Ordovician–Devonian carbonate facies and show optical properties similar to those of graptolites and vitrinite. They are generally rare in post-Permian strata, but have proved useful for thermal maturation studies of shallow marine deposits.
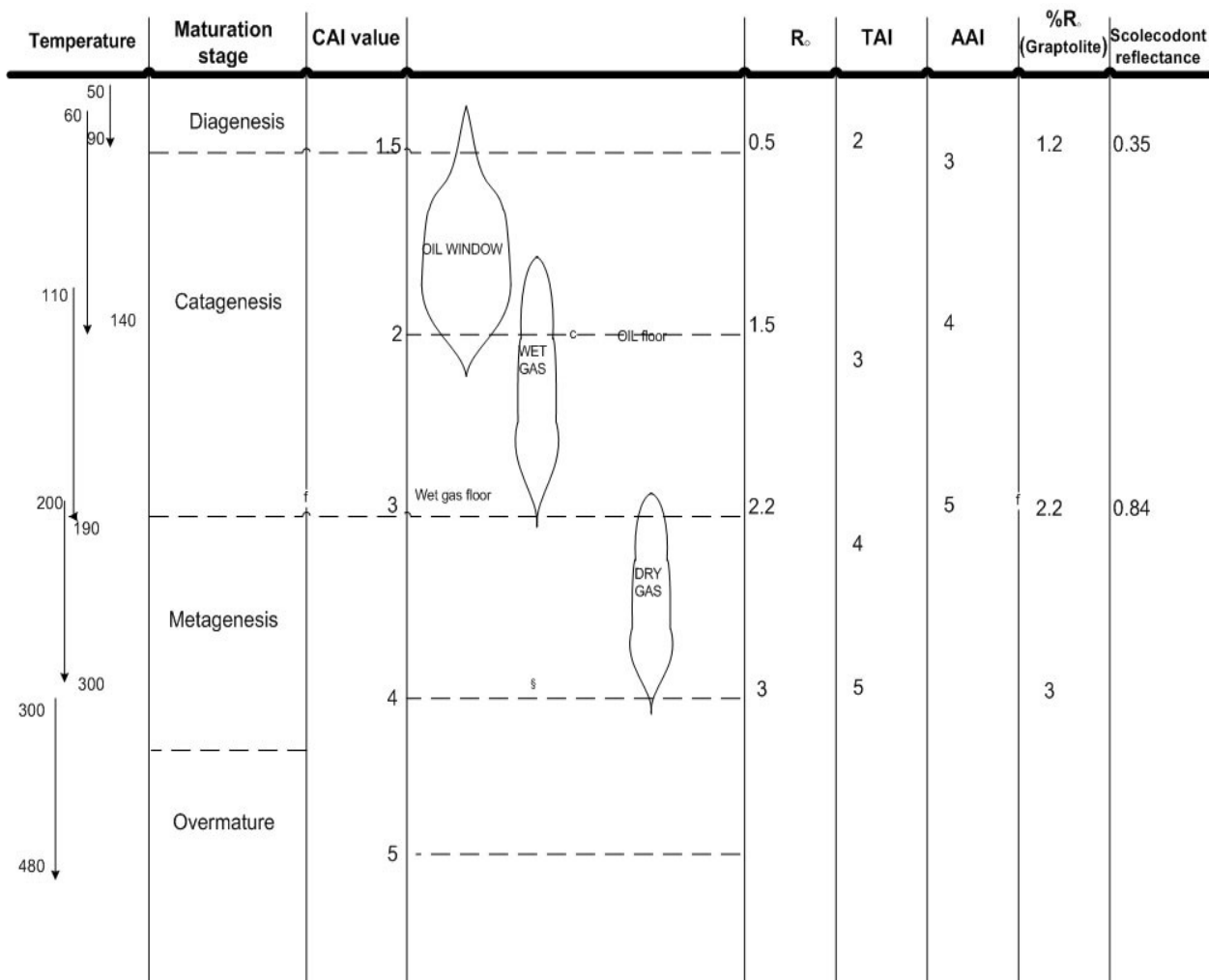Scolecodont reflectance values are similar to those of

Figure contents (left to right columns):

Temperature | Maturation stage | CAI value | [hydrocarbon zones] | R$_o$ | TAI | AAI | %R$_o$ (Graptolite) | Scolecodont reflectance

Temperature values: 50, 60, 90, 110, 140, 200, 190, 300, 300, 480

- Diagenesis — CAI 1.5 — R$_o$ 0.5 — TAI 2 — AAI 3 — %R$_o$ 1.2 — Scolecodont reflectance 0.35
- Catagenesis — OIL WINDOW, WET GAS — CAI 2, Oil floor — R$_o$ 1.5 — TAI 3 — AAI 4
- Wet gas floor — CAI 3 — R$_o$ 2.2 — TAI 4 — AAI 5 — %R$_o$ 2.2 — Scolecodont reflectance 0.84
- Metagenesis — DRY GAS — CAI 4 — R$_o$ 3 — TAI 5 — %R$_o$ 3
- Overmature — CAI 5

Fig. 19. Correlation between temperature, maturation stages, hydrocarbon zones and different thermal indicators. CAI, Conodont Alteration Index; R$_o$, vitrinite reflectance; TAI, Thermal Alteration Index; AAI, Acritarch Alteration Index.

graptolite reflectance in carbonate matrix, whereas graptolite reflectance in shale matrix shows higher values than the scolecodont reflectance (e.g. Goodarzi & Norford 1985; Goodarzi & Higgins 1987). This indicates that the organic material of scolecodonts has an higher hydrogen/carbon ratio than that of graptolites (Bertrand 1990). Scolecodonts with a reflectance between 0.35 and 0.84% corresponds to a CAI value of 1.5–2.5 and a vitrinite reflectance of 0.8–1.1, indicating that the sediments are within the oil window (e.g. Goodarzi & Higgins 1987).

As noted above, several groups of fossils can be used for thermal maturity studies. The results are, however, complicated by several factors such as time and diagenetic processes. It must also be emphasized that not all of the different groups reviewed in this paper are applicable throughout the geological column.

8 References

Anders, D.E. & Gerrild, P.M. 1984: Hydrocarbon generation in lacustrine rocks of Tertiary age, Unita Basin, Utah – Organic carbon, pyrolysis yield, and light hydrocarbons. *In* Woodward, J., Meissner, F.F. & Clayton, J.L. (eds.): *Hydrocarbon source rocks of the Greater Rocky Mountain Region*, 513–524. Denver Rocky Mountain Association of Geologists.

Armstrong, H.A. & Brasier, M.D. 2005: *Microfossils, second edition*, 296 pp. Blackwell Publishing Ltd, Oxford, UK.

Barnard, P.C., Cooper, B.C. & Fisher, M. 1976: Organic maturation and hydrocarbon generation in the Mesozoic sediments of the Sverdrup Basin,

Arctic Canada. 4th International Palynology Conference, Lucknow, India.

Belka, Z. 1991: Conodont color alteration patterns in Devonian rocks of the eastern Anti- Atlas, Morocco. *Journal of African Earth Sciences 12*, 417–428.

Bergström, S.M. 1980: Conodonts as paleotemperature tools in Ordovician rocks of the Caledonides and adjacent areas in Scandinavia and the British Isles. *Geologiska Föreningens i Stockholm Förhandlingar 102*, 337–392.

Bertrand, R. 1990: Correlations among the reflectances of vitrinite, chitinozoans, graptolites and scolecodonts. *Organic Geochemistry 15*, 565–574.

Bertrand, R. & Heroux, Y. 1987: Chitinozoan, graptolite, and scolecodonts reflectance as an alternative to vitrinite and pyrobitumen reflectance in Ordovician and Silurian strata, Anticosti Island, Quebec, Canada. *The American Association of Petroleum Geologists Bulletin 71*, 951–957.

Buiskool Toxopeus, J.M.A. 1983: Selection criteria for the use of vitrinite reflectance as a maturity tool. In Brooks, J. (eds): *petroleum geochemistry and exploration of Europe*, 295–307. Blackwell Scientific Publications, Oxford, UK.

Burrett, C.F. 1992: Conodont geothermometry in Paleozoic carbonate rocks of Tasmania and its economic implications. *Australian Journal of Earth Sciences 39*, 61–66.

Bustin, R.M., Orchard, M. & Mastalerz, M. 1992: Petrology and preliminary organic geochemistry of conodonts: implications for analyses of organic maturation. *International Journal of Coal Geology 21,* 231–282.

Butterfield, N.J. 2005: Probable Proterozoic fungi. *Paleobiology 31*, 165–182.

Cardott, B.J. & Kidwai, M.A. 1991: Graptolite reflectance as a potential thermal-maturation indicator. *Oklahoma Geological Survey Circular 92*, 203–209.

Carr, A.D. & Williamson, J.E. 1990: The relationship between aromaticity, vitrinite reflectance and maceral composition of coals: Implications for the use of vitrinite reflectance as a maturation parameter. *Organic Geochemistry 16,* 313–323.

Clarkson, E.N.K. 1998: *Invertebrate Palaeontology and Evolution, fourth edition,* 452 pp. Blackwell Science, Oxford, UK.

Claypool, G.E., Love, A.H. & Maughan, E.K. 1978:

Organic geochemistry, incipient metamorphism, and oil generation in black shale members of Phosphoria Formation, western interior United States. *AAPG Bulletin 62,* 98–120.

Craig, J.R., Vaughan, D.J. & Skinner, B.J. 2001: *Resources of the Earth – Origin, Use, and Environmental Impact, third edition*, 520 pp. Prentice Hall, New Jersey, USA.

Dow, W.G. 1977. Kerogen studies and geological interpretation. *Journal of Geochemical Exploration 7*, 79–99.

Dunoyer de Segonzac, G. 1970: The transformation of clay minerals during diagenesis and low-grade metamorphism: a review. *Sedimentology 15*, 281–346.

Durand, B. (ed.) 1980: *Kerogen insoluble organic matter from sedimentary rocks*, 519 pp. Èditions Technip, Paris, France.

El beialy, S.Y., Elatfy, H.S., Zavada, M.S., El khoriby, E.M. & Abu-Zied, R.H. 2010: Palynological, palynofacies, paleoenvironmental and organic geochemical studies on the Upper Cretaceous succession of the GPTSW-7 WELL, North Western Desert, Egypt. *Marine and Petroleum Geology 27*, 370–385.

Epstein, A.G., Epstein, J.B. & Harris, L.D. 1977: Conodont color alteration - an index to organic metamorphism. *U.S. Geological Survey Professional Paper 995*, 1–27.

Eriksson, M. 2000: Early Palaeozoic jawed polychaetes with focus on polychaetaspids and ramphoprionids from the Silurian of Gotland, Sweden. *Lund Publications in Geology 151*, 1–19.

Fermont, W.J.J. 1988: Possible causes of abnormal vitrinite reflectance value in paralic deposits of the Carboniferous in the Achterhoek area, The Netherland. *Organic Geochemistry 12,* 401–411.

Goodarzi, F. 1985: Dispersion of optical properties of graptolite epiderms with increased maturity in early Paleozoic organic sediments. *Fuel 64,* 1735–1740.

Goodarzi, F. & Higgins, A.C. 1987: Optical properties of scolecodonts and their use as indicators of thermal maturity. *Marine and Petroleum Geology 4,* 353–359.

Goodarzi, F. & Norford, B.S. 1985: Graptolite as indicators of the temperature histories of rocks. *Journal of the Geological Society London 142*, 1089–1099.

Goodarzi, F. & Norford, B.S. 1989: Variation of grap-

tolite reflectance with depth of burial. *International Journal of Coal Geology 11*, 127–141.

Goodarzi, F., Gentzis, T. & Harrison, C. 1992: The significance of graptolite reflectance in regional thermal maturity studies, Queen Elizabeth Islands, Arctic Canada. *Organic Geochemistry 18*, 347–357.

Gorter, J.D. 1984: Source potential of the Horn Valley Siltstone, Amadeus Basin. *Australian Petroleum Exploration Association Journal 24*, 225–236

Gradstein, F., Ogg, J. & Smith, A. 2004: *A Geologic Time Scale 2004,* 589 pp. Cambridge University Press, Cambridge, UK.

Gray, J. & Boucot, A.J. 1975: Color change in pollen and spores: A Review. *Geological Society of America Bulletin 86*, 1019–1033.

Harley, J., 1891: The Ludlow bone-bed and its crustacean remains. *Quarterly journal of the Geological society of London 17*, 542–553.

Huang, W. 1996: Experimental study of vitrinite maturation effects of temperature, time, pressure, water, and hydrogen index. *Organic Geochemistry 24,* 233–241.

Hunt, J.M. 1984: Generation and migration of light hydrocarbons. *Science 226,* 1265–1270.

Hunt, J.M. 1995: *Petroleum Geochemistry and Geology, Second edition*, 743 pp. W.H. Freeman and Company, New York.

Hyde, H.A.& Williams, D.A. 1944: The right word. *Pollen Analysis Circular 8*, 6.

Ismail, M. & Shamsuddin, A.H.M. 1991: Organic matter maturity and its relation to time, temperature and depth in the Bengal foredeep, Bangladesh. *Journal of Southeast Asian Earth Sciences 5*, 381–390.

Jarzen, D.M. & Nichols, D.J. 1996: Chapter 9: Pollen. *In* Jansonius, J. & McGregor, D.C. (eds): *Palynology: Principles and Applications 1*, 261–291. American Association of Stratigraphic Palynologists Foundation 1.

Kurylowicz, L.E., Ozimic, S., Mckirdy, D.M., Kantsler, A.J. & Cook, A.C. 1976: Reservoir and source rock potential of the Larapinta Group, Amadeus basin, central Australia. *Australian Petroleum Exploration Association Journal 16,* 49–65.

Legall, F.D & Barnes, C.R. 1980: Application of conodont and palynomorph color alteration studies to thermal maturation history, southern Ontario. The *American Association of Petroleum Geologists Bulletin 64, 5,* 738–739.

Legall, F.D., Barnes, C.R. & Macqueen, R.W. 1981: Thermal maturation, burial history and hotspot development, Paleozoic strata of southern Ontario -Quebec, from conodont and acritarch colour alteration studies. *Bulletin of Canadian Petroleum Geology 29,* 492–539.

Le Tran, K. 1972: Geochemical study of hydrogen sulfide absorbed in sediment. *In* von Gaertner, H.R., & Wehner, H. (eds.): *Advances in organic geochemistry 1971*, 717–726. Pergamon Press, Oxford.

Link, C.M., Bustin, R.M. & Goodarzi, F. 1990: Petrology of graptolites and their utility as indices of thermal maturity in Lower Palaeozoic strata in northern Yukon, Canada. *International Journal of Coal Geology 15*, 113–135.

Marshall, J.E.A. 1990: Determination of thermal maturity. *In* Briggs, D.E.G. & Crowther, P. (eds): *Palaeobiology – a synthesis*, 511–515. Blackwell Scientific Publications, Oxford, UK.

Marshall, J.E.A.1991. Quantitative spore colour. *Journal of the Geological Society, London 148*, 223–233.

Marshall, C.P., Rose, H.R., Lee, G.S.H., Mar, G.L. & Willson, M.A. 1999: Structure of organic matter in conodonts with different colour alteration indexes. *Organic Geochemistry 30*, 1339–1352.

Milson, C. & Rigby, S. 2010: *Fossils at a Glance, second edition,* 159 pp. Wiley-Blackwell, Chichester, West Sussex, UK.

Mukhopadhyay, P.K. 1994: Chapter 1: Vitrinite reflectance as maturity parameter – Petrography and molecular characterization and its application to basin modelling. *In* Mukhopadhyay, P.K., Dow, W.G. (eds.): *Vitrinite Reflectance as a Maturity Parameter, Application and Limitations*, 1–24. American Chemical Society, Washington DC.

Olsson-Borell, I. 2003: Thermal history of the Phanerozoic sedimentary successions of Skåne, southern Sweden, and implications for applied geology. *Litholund theses 3*, 1–23.

Pietzner, H., Vahl, J., Werner, H. & Ziegler, W. 1968: Zur chemischen Zusammensetzung und Mikromorphologie der Conodonten. *Palaeontographica Abteilung A128*, 115–152.

Playford. G. & Dettmann, M. E. 1996: Chapter 8: Spores. *In* Jansonius, J. & Mcgregor, D.C. (eds): *Palynology: Principles and Applications 1*, 227–260. American Association of Stratigraphic Palynologists Foundation 1.

Rejebian, V.A., Harris, A.G. & Huebner, J.S. 1987:

Conodont color and textural alteration: An index to regional metamorphism, contact metamorphism, and hydrothermal alteration. *Geological Society of America Bulletin 99*, 471–479.

Rubinstein, C.V., Gerrienne, P., de la Puente, G.S., Astini, R.A. & Steemans, P. 2010: Early Middle Ordovician evidence for land plants in Argentina (eastern Gondwana). *New phytologist 188*, 365–369.

Selley, R.C. 1985: *Elements of Petroleum Geology*, 449 pp. W.H. Freeman and Company, New York.

Staplin, F.L. 1969: Sedimentary organic matter, organic metamorphism, and oil and gas occurrence. *Bulletin of Canadian Petroleum Geology 17*, 47–66.

Strother, P.K. 1996: Chapter 5: Acritarchs. *In* Jansonius, J. & McGregor, D.C. (eds): *Palynology: Principles and Applications 1*, 81–106. American Association of Stratigraphic Palynologists Foundation 1.

Sweeney, J.J., Burnham, A.K. & Braun, R.L. 1987: A model of hydrocarbon generation from type I kerogen: application to Uinta Basin, Utah. *The American Association of Petroleum Geologists Bulletin 71*, 967–985.

Sweet, W.C. 1988: The Conodonta: *Morphology, Taxonomy, Paleoecology, and Evolutionary History of a Long-Extinct Animal Phylum*, 212 pp. Oxford Monographs on Geology and Geophysics 10. Oxford University Press, Oxford.

Symons, D.T.A. & Cioppa, M.T. 2002: Conodont CAI and magnetic mineral unblocking temperatures: implications for the Western Canada Sedimentary Basin. *Physics and chemistry of the Earth 27,* 1189–1193.

Talyzina, N.M. & Moczydłowska, M. 2000: Morphological and ultrastructural studies of some acritarchs from the Lower Cambrian Lükati Formation, Estonia. *Review of Palaeobotany and Palynology 112*, 1–21.

Taylor, T.N. 1996: Chapter 14: In situ pollen and spores in plant evolution introduction. *In* Jansonius, J. & McGregor, D.C. (eds): *Palynology: Principles and Applications 1*, 389–390. American Association of Stratigraphic Palynologists Foundation 1.

Tissot, B.P. & Welte, D.H. 1978: *Petroleum Formation and Occurrence – A New Approach to Oil and Gas Exploration*, 538 pp. Springer-Verlag, Berlin, Heidelberg, New York.

Travers, A. 1988: *Paleopalynology*, 600 pp. Unwin Hyman, London.

Tucker, M.E. 2001: *Sedimentary petrology: an introduction to the origin of sedimentary rocks, third edition,* 262 pp. Blackwell Science, Oxford, UK.

Teichmüller, M. 1958: Métamorphisme du charbon et prospection du pétrole. *Revue de l'industrie minerale, Special Issue,* 1–15.

Ujiié, Y. 2001: Brightness of pollen as an indicator of thermal alteration by means of a computer-driven image processor: statistical thermal alteration index (stTAI). *Organic Geochemistry 32,* 127–141.

Utting, J. & Hamblin, A.P. 1991: Thermal maturity of the Lower Carboniferous Horten Group, Nova Scotia. *International Journal of Coal Geology 13,* 439–456.

Vecoli, M. & Hérissé, A.L. 2004: Biostratigraphy, taxonomic diversity and patterns of morphological evolution of Ordovician acritarchs (organic-walled microphytoplankton) from the northern Gondwana margin in relation to palaeoclimatic and palaeogeographic changes. *Earth-Science Reviews 67,* 267– 311.

Voldman, G.G., Albanesi, G.L. & Campo, M.D. 2008: Conodont palaeothermometry of contact metamorphism in Middle Ordovician rocks from the Precordillera of western Argentina. *Geological Magazine 145,* 449–462.

Williams, H.S., Burden. E.T. & Mukhopadhyay, P.K. 1998: Thermal maturity and burial history of Paleozoic rocks in western Newfoundland. *Canadian Journal of Earth Sciences 35*, 1307–1322.

Willman, S. & Moczydłowska, M. 2007: Wall ultrastructure of an Ediacaran acritarch from the Officer Basin, Australia. *Lethaia 40*, 111–123.

Yin, H.F., Zhang, K.X., Tong, J.N., Yang, Z.Y. & Wu, S.B. 2001: The Global Stratotype Section and Point (GSSP) of the Permian-Triassic boundary. *Episodes 24*, 102–114.

Appendix I - Experimentally Induced Color Alteration

Introduction

Conodonts progressively undergo color change from pale yellow through brown to black as a result to increasing temperature through time. This conodont colour change is attributed to loss of organic matter, release of water of crystallization and recrystallization (Epstein et al. 1977; Rejebian et al. 1987). The changes in conodont colour can be detected in a laboratory by different methods such as Open-Air Heating Runs technique with water or without water. This report provides experimental data for the use of conodonts as thermal indicators by applying an Open-Air Heating Runs technique.

Method

For consistency with Epstein et al. (1977), the Open-Air Heating Technique without water was used herein to detect the conodont colour change because this method is the most useful as it can show the colour change in conodonts already after half an hour if heated to 500°C (Epstein et al. 1977). The conodonts were heated in an electric furnace to a temperature of 500 °C. For the purpose, eighteen specimens of thermally unaltered (CAI 1) conodont elements from the 'Täljsten' interval in the Middle Ordovician Holen Limestone of Västergötland, Sweden, were put in the oven. These belonged to the species *Baltonious norrlandicus, Periodon macrodentata, and the coniform Drepanodus arcuatus*. Subsequently three at the time were taken out after the schedule outlined below. The laboratory work was carried out as follows:

Before carrying out the experiment, the oven was tested for its stability by measuring its temperature. The oven was set to 500 degrees and its temperature was tested by a thermometer every 15 minutes.

1- Six glass containers with three conodonts each were prepared.

2- All samples were placed in the oven and heated to 500°C and subsequently the beakers were taken out at the time interval listed below:

A- First sample heated during one hour

B- Second sample heated during two hours

C-Third sample heated during three hours

D- Fourth sample heated during four hours

E- Fifth sample heated during eight hours

F- Sixth sample heated during 24 hours.

3- Finally photographs were taken in a xxx microscope of all conodonts and compared with standard sample.

Result

These experimental studies on conodonts with specimens heated to 500 °C during different time intervals reveal that samples that were heated for a long time get a darker colour than those heated during a short time (Fig. 2). This colour change expressed as a response to time can be listed as below:

1- The sample that was heated during one hour has a colour range from pale yellow to pale brown and has CAI=1 (Fig. 1)

2- The sample that was heated during two hours has a colour range from pale brown to very pale brown and has a CAI range from 1 to 1.5 (Fig. 2).

3- The sample that was heated during 3 hours has a brown color and CAI=2 (Fig. 3).

4- The sample that was heated during 4 hours is characterized by a colour change from brown to dark brown and the conodont surface changes from smooth to pitted. CAI value ≤ 2 >3 (Fig. 4).

5- The sample that was heated during 8 hours is characterized by a colour change from dark brown to very dark grayish brown and it has a white or crystal colour in some places. CAI value =3 (Fig. 5)

6- The sample that was heated during 24 hours shows a brown colour. The conodonts in this sample show uneven colors, darker in some areas while lighter brown in other areas. Specimens in this sample should be darker or has a black colour compare to last sample (Fig. 6).

A small variation in the degree of heating took place during the heating process while opening and closing the oven, when the temperature goes down from 499°C to 450°C after opening the oven to take out the

Fig. 1. Middle Ordovician conodonts heated at 500°C during one hour. Colour range from pale yellow to pale brown, CAI=1.

Fig. 2. Middle Ordovician conodonts heated to 500°C during two hours. Colour ranges from pale brown to very pale brown. CAI = 1 to 1.5.

Fig. 3. Middle Ordovician conodont heated to 500°C during three hours. It has a brown color. CAI value = 2.

Fig. 4. Middle Ordovician conodont heated to 500°C during four hours. It has a brown to dark brown color. CAI

Fig. 5. Middle Ordovician conodont heated to 500°C during 8 hours. It has a dark brown to very dark grayish brown col-

Fig. 6. Middle Ordovician conodont heated to 500°C during 24 hours. It has a brown to dark brown color.

Table.1. show the change in temperature during the experimental work, including a mean temperature and standard

Time	Temperature °C
9:30	503
9:45	499
10:00	499
10:15	499
10:30	453
10:45	499
11:00	499
11:15	499
11:30	455
11:45	499
12:00	499
12:15	499
12:30	450
12:45	499
13:00	499
13:15	499
13:30	450
Mean	488
Standard deviation	20.11

Fig. 7. Showing the temperature variation during time.

sample. It then increases to 521 °C after closing the oven (see Table 1; Fig. 7). The increase in temperature continues during five minutes before the temperature returns to 499°C. However the mean temperature equal to 488°C and the Standard deviation is 20.11 which means that there are small variations in the degree of heating.
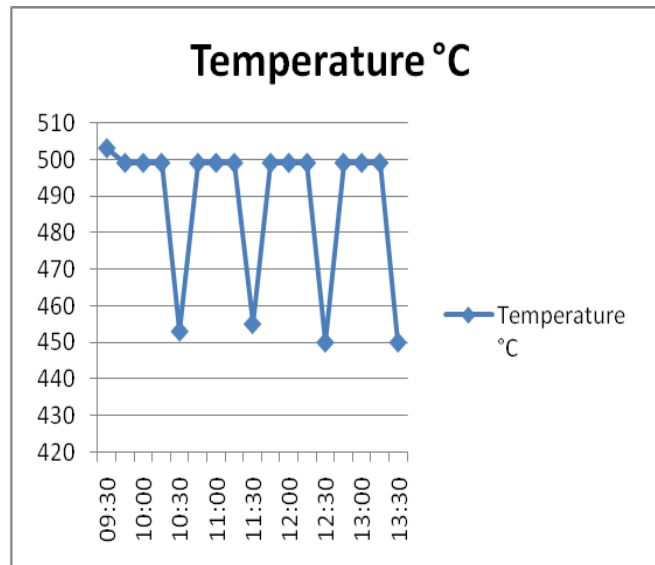
Discussion

The analysis shows that conodonts that were heated a longer time get a darker colour in comparison to those heated during a short time. This is due to the succes-

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sive loss of organic matter and recrystallization. When the results are compared with the study by Epstein et al. (1977; fig. 5) similar results are seen; i.e. they show similar colour change from pale yellow through brown to black as a function to increasing temperature and longer time. There are several variations to this theme, however. For example, in contrast to the variations showed in the samples that were analysed by Epstein et al. (1977) the studied material shows less change through the 24 hours. Another difference is that some of the studied specimens show a variation in colour within the same conodont specimen. This can possibly be attributed to factors such as shape and size of specimen. One of the most important differences is found in the sample that was heated during 24 hours and which shows brown colour although it was heated a long time compared to the other samples. This sample should have reflected a darker or black colour. The causes for this variation are unknown but it can be attributed to robustness or delicateness of conodont specimen (Epstein et al. 1977; Rejebian et al. 1987).

Conclusions

1- The conodonts color alteration index is an important tool in assessing thermal maturity and organic metamorphism because it shows a colour change when the conodonts are heated. This colour change indicates the stage of metamorphism.

2- The CAI = 1.5 can be detected by Open-Air Heating Runs method by heating conodont for 500°C during one to two hours.

3- CAI value =2 is detected by Open-Air Heating Runs method when heating conodonts to 500°C during three hours.

4- CAI value= 3 is reached when the conodont for 500°C to and time range from four to eight hours.

5- To get high CAI value it needs to heating sample more than 500°C and long time.

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