

Jurassic and Cretaceous clays of the northern and central North Sea hydrocarbon reservoirs reviewed

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ABSTRACT: The principal clays of the northern and central North Sea are illite (sometimes with interlayered smectite) and kaolin. Chlorite is only locally important. While it has been proposed that kaolin within North Sea sandstones is detrital in origin, the majority of workers have concluded that it is authigenic, largely the product of feldspar alteration. Kaolin is found within a wide range of sedimentary settings (and within shales) apparently defying the notion that kaolin is an indicator of meteoric water flushing. Within sandstones, the earliest authigenic kaolin has a vermiform morphology, the distribution of which is controlled by the availability of detrital mica to act as a nucleus, and the composition of the post-depositional porewaters. This vermiform kaolin formed in meteoric water, the presence of which is easily accounted for below sub-aerial exposure surfaces in non-marine formations, and below unconformities over marine units. In fully marine sands, and even marine shale units, kaolin still occurs. It has therefore been suggested that even these locations have been flushed with meteoric water.

Early vermiform kaolin recrystallizes to a more blocky morphology as burial proceeds, at least in the Brent Group. Blocky kaolin has been reported as growing before, synchronously with, and after the formation of quartz overgrowths, though oxygen isotope studies support low-temperature growth, pre-quartz. Blocky kaolin may form during meteoric flushing associated with lower Cretaceous uplift and erosion, though it is found in fault blocks that are thought to have remained below sea level. Here, the kaolin may form in stagnant meteoric water, relics of the post-depositional porewater. It has also been proposed that the blocky kaolin grew in ascending basinal waters charged with carboxylic acids and CO₂, though this hypothesis is not supported by stable oxygen isotope data. Some of the blocky kaolin is dickite, the stable polymorph above ~100°C.

Illite occurs almost ubiquitously within the clastic sediments of the North Sea. An early pore-lining phase has been interpreted as both infiltrated clastic clay, and as an early diagenetic phase. Early clays may have been quite smectite-rich illites, or even discrete smectites. Later, fibrous illite is undoubtedly neoformed, and can degrade reservoir quality significantly. Both within sandstones and shales, there is an apparent increase in the K content over 4 km of burial, which could be due to dilution of the early smectite-rich phase by new growth illite, or to the progressive illitization of existing I-S. Much of the 'illite' that has been dated by the K-Ar method may therefore actually be I-S.

The factors that control the formation of illite are only poorly known, though temperature must play a role. Illite growth has been proposed for almost the entire range of diagenetic temperatures (e.g. 15–20°C, Brent Group; 35–40°C, Oxfordian Sand, Inner Moray Firth; 50–90°C, Brae formation; 100–110°C, Brent Group; 130–140°C, Haltenbanken). It seems unlikely that there is a threshold temperature below which illite growth is impossible (or too slow to be significant), though this is a recurring hypothesis in the literature. Instead, illite growth seems to be an event, commonly triggered by oil emplacement or another change in the physiochemical conditions within the

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sandstone, such as an episode of overpressure release. Hence illite can grow at any temperature encountered during diagenesis.

Although there is an extensive dataset of K-Ar ages of authigenic illites from the Jurassic of the North Sea, there is no consensus as to whether the data are meaningful, or whether the purified illite samples prepared for analysis are so contaminated with detrital phases as to render the age data meaningless. At present it is unclear about how to resolve this problem, though there is some indication that chemical micro-analysis could help. It is a common belief that illite ages record the timing of oil charge, and so can be used to calibrate basin models.

Grain-coating Fe-rich chlorite cements can preserve exceptional porosity during burial. They are found in marginal marine sandstones, formed during diagenesis from precursor Fe-rich clays such as berthierine or verdine.

KEYWORDS: Cretaceous, Jurassic, clays, North Sea, hydrocarbon reservoirs.

The Jurassic and Cretaceous strata of the North Sea host the majority of the hydrocarbons within the UK Continental Shelf? (UKCS), including both reservoirs and source rocks. The Middle Jurassic Brent Group is the principal reservoir unit of the Northern North Sea, while the Upper Jurassic Fulmar Formation is the most important target in the Central North Sea (Fig. 1). The majority of studies on these units has been by, or on behalf of, the oil companies, and only a (small?) proportion of the results are in the public domain. The majority of the work has focused on reservoir sandstones, with interbedded shales regarded merely as barriers to fluid flow and potential reservoir seals. Clay

minerals are present within the sandstones both as detrital minerals and as new growth, authigenic minerals. The latter have been extensively studied as even a small percentage of fibrous, authigenic illite in a sandstone can dramatically reduce permeability, and hence the rate at which oil can be extracted commercially. For example, the permeability of the Etive, Ness and Tarbert Formations is independent of depth above 3.1 km, but decreases linearly below this, attributed by Giles *et al.* (1992) as due to illite precipitation and related modifications to the pore network.

This review is arranged by stratigraphy, starting with the Lower Jurassic. The vast majority of the work is concerned with sandstones, though shales are also reviewed. The most important clay minerals are kaolin and illite or illite-smectite, with chlorite only common locally. Illite and kaolin are described separately within each stratigraphic section, despite the observation that illite is commonly formed at the expense of kaolin.

LOWER JURASSIC – NANSEN, STATFJORD, TOFTE AND TILJE FORMATIONS

There are few published studies of the Lower Jurassic, as there are relatively few commercial reservoirs within this interval. The Statfjord Formation spans the Triassic–Jurassic boundary, but is included here. Authigenic kaolin occurs both as verms and blocky crystals, the significance of which is discussed below. There is no positive evidence for the presence of the kaolin polymorph dickite (Purvis, 1995). Illite and quartz cements become abundant below 3–4000 m, unless inhibited by early chlorite cements (Ramm and Ryseth, 1996). Grain rimming Fe-chlorite cements also

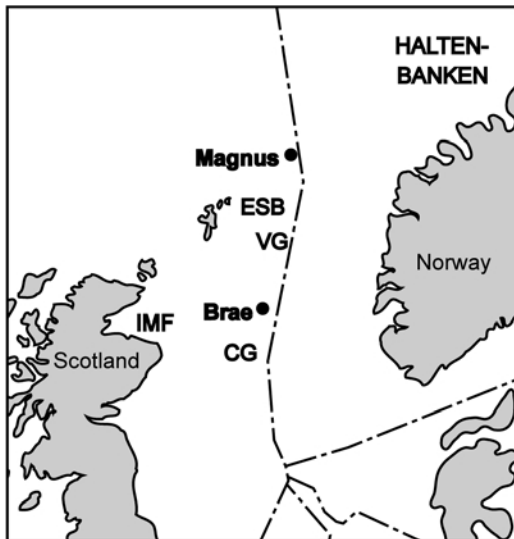


FIG. 1. Location map. ESB is East Shetland Basin, VG is Viking Graben, IMF is Inner Moray Firth and CG is Central Graben. The dot-dash lines show the boundaries between the national sectors.

occur in the marginal marine Nansen, Tofte and Tilje Formations (Purvis, 1995; Ehrenberg *et al.*, 1998). These cements are associated with very high-porosity sands, described for the Middle Jurassic below. In the Nansen Formation, the chlorite cements apparently pre-date the early dolomite and kaolin cements (Purvis, 1995), in contrast to the very similar chlorite cements of the Middle-Jurassic sands that post-date carbonate cementation, see below.

MIDDLE JURASSIC – BRENT GROUP, SLEIPNER AND GARN FORMATIONS

This section is, inevitably, dominated by studies of the Brent Group, though valuable contributions have been made by workers looking at the Norwegian Garn Formation.

Middle-Jurassic kaolin

Because of the very fine-grained nature of clay cements, even early studies of the Brent Group sandstones utilized scanning electron microscopy (SEM) for the determination of petrographic relationships (e.g. Hancock & Taylor, 1978; Sommer, 1978), while some more modern studies have benefited from literally hundreds of hours of SEM observation (e.g. Osborne *et al.*, 1994). The usefulness of early studies is limited by the lack of modern stratigraphy, and by the fact that, for commercial reasons, wells and fields were not identified. All recent petrographic studies of the Brent Group have recorded at least two phases of kaolin growth: early vermiform and later blocky kaolin. Some have also distinguished kaolin associated with detrital mica as a third phase, though some regard this as either synchronous with the formation of, or equivalent to, vermiform kaolin (Kantorowicz, 1984; Lønøy *et al.*, 1986; Thomas, 1986; Glasmann *et al.*, 1989c; Giles *et al.*, 1992; Harris, 1992; Hogg *et al.*, 1993; Osborne *et al.*, 1994; Hassouta *et al.*, 1999; Girard *et al.*, 2002; Wilkinson *et al.*, 2004). Hurst & Irwin (1982) proposed a model for kaolin morphology whereby vermiform kaolin grew in meteoric porewaters, and blocky ('euhedral') kaolin grew in marine porewaters. This model is rejected here, as discussed below.

There is some controversy concerning the relative timing of the formation of kaolin, which is

important as it has implications for the origin of the porefluids from which growth occurred. In one of the first published diagenetic studies of the Brent Group, Sommer (1978) identified kaolin growth as post-dating the formation of quartz overgrowths, as did Blackburn (1984). Other authors have deduced that the formation of quartz overgrowths and kaolin were synchronous (Lønøy *et al.*, 1986; Thomas, 1986; Scotchman *et al.*, 1989; Giles *et al.*, 1992; Haszeldine *et al.*, 1992; McAulay *et al.*, 1994; Girard *et al.*, 2002). Sommer (1978) observed the penetration of kaolin booklets into quartz overgrowths, and interpreted this to indicate that the quartz overgrowths had been deeply corroded during the growth of the kaolin. This meant that the formation of quartz overgrowths had to pre-date the lower Cretaceous uplift, when Sommer (1978) postulated that meteoric water descended from the resulting unconformity and the kaolin grew. The same kaolin-quartz overgrowth texture has been observed in other studies, but has been interpreted as kaolin formation prior to the growth of the quartz overgrowths that have enclosed pre-existing kaolin (Blanche & Whitaker, 1978; Glasmann *et al.*, 1989c; Harris, 1992; Hogg *et al.*, 1993; Bjørlykke *et al.*, 1992; Osborne *et al.*, 1994; Wilkinson *et al.*, 2004). In general, petrographic textures of intergrowth and corrosion can be very difficult to distinguish, especially where it is suspected that one mineral has grown corrosively into a pre-existing phase. Hurst & Irwin (1982) summarized the contemporary state of knowledge concerning diagenesis and concluded that the relationship between quartz overgrowths and kaolin is controlled by depositional setting – in fluvial sediments, kaolin precipitates first, but this is reversed in more marine settings. They related this to the probability of the sediments undergoing an early meteoric water flush-which is very probable in fluvial systems but much less likely for fully marine sediments. Here we conclude that the growth of both early vermiform and late blocky kaolin pre-dates the majority of quartz overgrowths, with most kaolin forming below 80°C and most quartz overgrowths forming above this temperature. This is probably related to the deltaic setting of the Brent Group, where the majority of the marine sediments were overlain by the non-marine Ness formation, when meteoric water could have penetrated into the subsurface. A fall in relative sea level in post-Ness times may have enhanced subsequent meteoric water penetration (McAulay *et al.*, 1994).

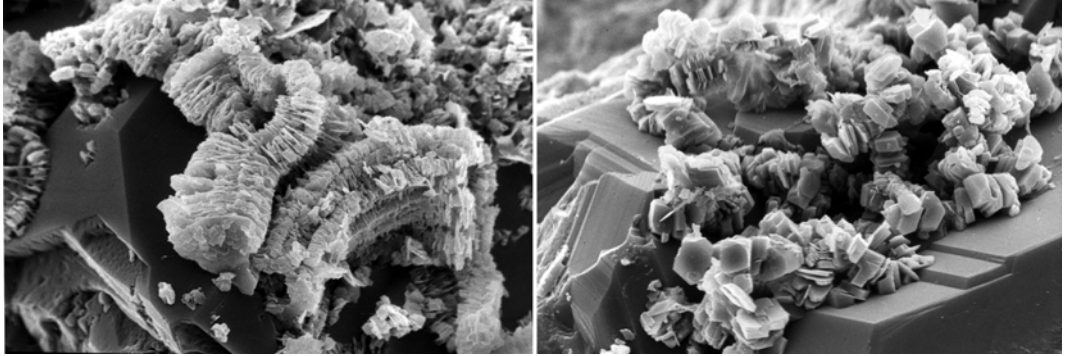


FIG. 2. Kaolin morphologies from the Brent Group, Cormorant field. Left: vermiform kaolin from well 211/21-CN40 at 3466 m drillers' depth (image 55 μm wide). Right: blocky kaolin from well 211/21-CN23 at 4401 m drillers depth (image 110 μm wide).

Early authigenic kaolin. The vermiform crystal aggregates (verms) and kaolin associated with detrital mica are formed during early diagenesis, as they are found enclosed within early calcite concretions (Fig. 2; Glasmann *et al.*, 1989c; Osborne *et al.*, 1994; Wilkinson *et al.*, 2004). Authigenic kaolin verms are relatively large (up to 200–300 μm maximum dimension) and can be observed optically in thin section. Osborne *et al.* (1994) claimed to be able to measure quantitatively the abundances of the vermiform and blocky kaolin using optical thin-section point-counting, though given that the two polymorphs are frequently intergrown, this seems optimistic. Despite the potential problems with the method, a compilation of point-count data showed a progressive change from vermiform to blocky kaolin between, 2000 and 4300 m burial depth (Osborne *et al.*, 1994; Fig. 3). This dataset is evidence that the blocky kaolin progressively replaces the vermiform kaolin during burial, as observed petrographically.

Thomas (1986) placed vermiform kaolin before kaolin intergrown with detrital mica in the paragenetic sequence. Greenwood *et al.* (1994) reversed the order of these two phases in the Sleipner Formation of the Brae area. They did not recognize a transformation from vermiform to blocky kaolin, unlike most other workers. Greenwood *et al.* (1994) observed identical kaolin morphologies in both sandstones and mudrocks. Osborne *et al.* (1994) suggested that the vermiform crystals and those associated with detrital mica shared a common origin. Kaolin is apparently more common in the micaceous Rannoch and Ness Formations than in the more mica-deficient parts

of the Brent Group (Harris, 1992), and there is a good correlation between detrital mica content and kaolin in the Statfjord Field (Bjørlykke & Brendsall, 1986). Together, these observations could be taken to suggest that every kaolin verm has nucleated on detrital mica, even if the mica nucleus is not now readily apparent. This would explain the unusual crystal morphology which is

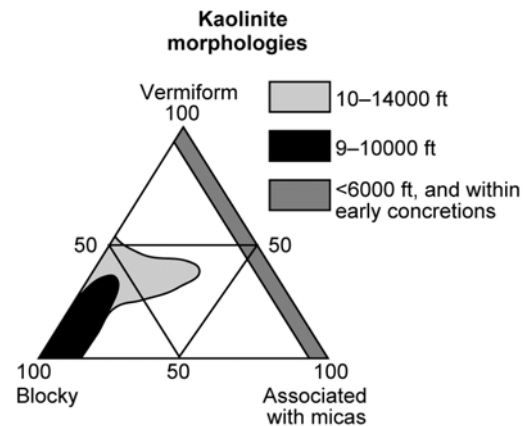


FIG. 3. Point-count data showing the variation in kaolin morphology with depth of burial (redrawn from Osborne *et al.*, 1994). As burial depth increases, the proportion of blocky kaolin increases. In very shallow samples, there is only vermiform kaolin and kaolin associated with the alteration of detrital muscovite. The same assemblage is preserved within calcite concretions, regardless of present-day burial depth. Note that some workers do not distinguish between vermiform kaolin and kaolin associated with mica, and it may be that all vermiform kaolin is actually nucleated upon detrital mica.

barely stable under the conditions of formation (given that recrystallization to the blocky morphology begins at only slightly higher temperatures than those at which precipitation took place). Giles *et al.* (1992) observed the most kaolin in the Etive Formation (and not the more micaceous Rannoch and Ness Formations). However, this may be related to the depths for which samples were available: Bjørlykke *et al.* (1992) presented a similar regional dataset in which most of the relatively shallow samples are from the Etive Formation, and as there is a regional decrease in kaolin abundance with depth (due to illitization), so the Etive samples are the most kaolin-rich. Giles *et al.* (1992) observed a facies control on kaolin abundance within the Ness Formation: cross-bedded and massive sands were more kaolin-rich. This pattern was not observed within the other formations, and has not been reported in other studies.

The majority of authors have assumed that since kaolin is formed in acidic groundwater at relatively low temperatures (e.g. Fairbridge, 1967) then this equates to meteoric water (e.g. Glasmann *et al.*, 1989c; Harris, 1992; Bjørkum *et al.*, 1990; Giles *et al.*, 1992). Isotopic studies suggest growth of vermiform kaolin at temperatures of <40 or 50°C (Wilkinson *et al.*, 2004; Osborne *et al.*, 1994). Meteoric water could have penetrated the Brent Group either immediately following deposition, or during the formation of the regional lower Cretaceous unconformity. Petrographic evidence suggests that the vermiform kaolin is formed very early, at least before some of the calcite concretions within which it is enclosed (Osborne *et al.*, 1994; Wilkinson *et al.*, 2004). It has therefore been assumed that the vermiform kaolin formed soon after the deposition of the Brent Group (McAulay *et al.*, 1994), when meteoric waters collected from the surface of the delta formed a meteoric water lens extending out below the sea (Kantorowicz, 1984). This is supported by the available isotopic data (Fig. 4). The dataset of Osborne *et al.* (1994) is derived by analysis of separates of pure vermiform kaolin, as this is the only morphology in the Emerald field. This shows a range of $\delta^{18}\text{O} = 16.1\text{--}18.5\text{‰}$ SMOW, which is consistent with precipitation from meteoric porewaters at ~20–30°C (Fig. 4).

There is general agreement that the early (vermiform) kaolin was sourced by the dissolution of detrital feldspar, most likely K-feldspar (Blanche & Whitaker, 1978; Glasmann *et al.*, 1989C; Giles *et*

al., 1992; Harris, 1992; Osborne *et al.*, 1994). There is abundant petrographic evidence of K-feldspar dissolution in the form of partly-dissolved relics and grain outlines preserved as authigenic overgrowths or clay rims. Some of these can be seen preserved within early calcite concretions, which proves that at least some K-feldspar dissolution took place during early diagenesis. Osborne *et al.* (1994) concluded that the K-feldspar dissolution reaction was isochemical for silica and aluminium, though no mention was made of the fate of the K. The chemical reaction used by Osborne *et al.* (1994) for their mass-balance calculation was based on an equation proposed by Giles & de Boer (1990), and this merely shows the K going into solution. Unless the K is removed, the reaction will halt. Either the K is removed in flowing groundwater, or is absorbed by clay mineral reactions either within the sandstones or in adjacent shales. Blanche & Whitaker (1978) had earlier suggested that detrital 'degraded illite' was essential as a K acceptor, and that this clay mineral was a product of weathering.

It seems likely that the vermiform kaolin is invariably the polymorph kaolinite, and not the higher-temperature polymorph dickite. Hancock & Taylor (1978) refer only to kaolinite, and their SEM images show only vermiform kaolin. Although the field on which the study is based is not identified, and burial depths are not stated, burial must be quite deep (~4000 m) as feldspar has apparently all dissolved and illitization of kaolinite is more-or-less complete, at least in the lower sections of the field. This suggests that, in this example, kaolinite is transforming to illite without recrystallizing into the blocky kaolin crystals observed elsewhere (see below), and probably without transforming to dickite.

Blocky kaolin. The blocky kaolin is observed to post-date the early vermiform kaolin, which it progressively replaces during burial (Figs 2, 3). Using stable oxygen isotope ratios, Wilkinson *et al.* (2004) estimated that the blocky kaolin grew at 45–70°C, via a dissolution-reprecipitation mechanism from the vermiform crystals. This formation temperature is close to other estimates of the temperature of formation of blocky kaolin: 30–80°C of Osborne *et al.* (1994) in a multi-field isotopic study of kaolin in the Brent Group, and 45–60°C of Glasmann *et al.* (1989C) from the Heather field. Giles *et al.* (1992) and Girard *et al.* (2002) derived significantly higher temperatures for

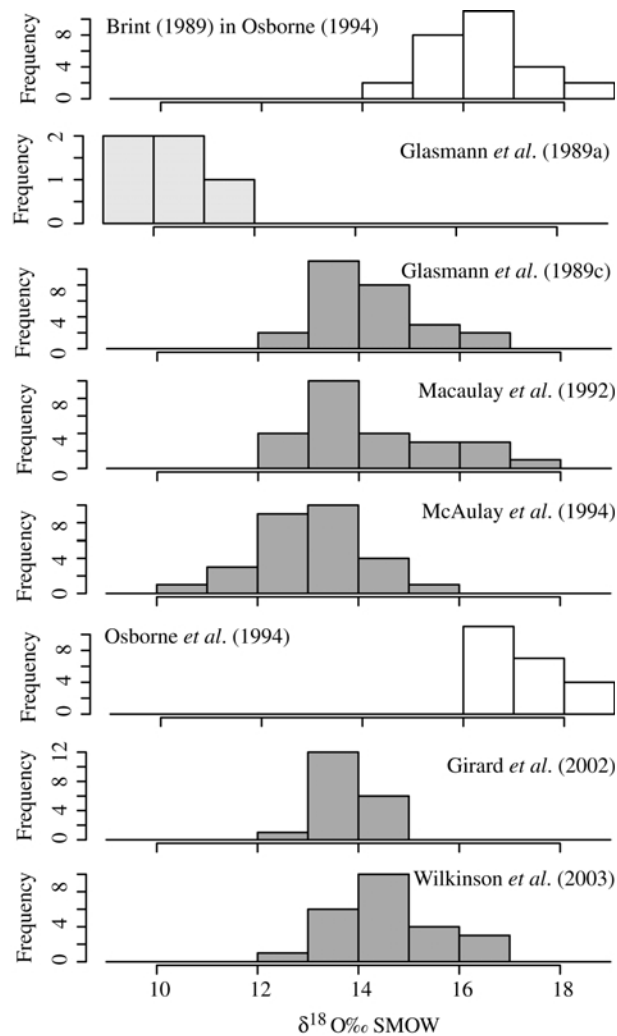


FIG. 4. Histograms of published stable oxygen isotope ratios of kaolin ($\delta^{18}\text{O}\text{‰ SMOW}$) from the Brent Group and the Magnus sandstone, northern North Sea. The data set of Osborne *et al.* (1994) is early, vermiform kaolin, while that of Brint (1989) is mixed vermiform and blocky kaolin (white on Figure). The remainder are at least predominantly blocky kaolin. All the blocky kaolin data sets have a similar mean and range, apart from the data of Glasmann *et al.* (1989a). Here the tabulated data (as plotted) is at odds with the range of data stated in the text of Glasmann *et al.* (1989a) and may be in error.

blocky kaolin growth of 90–100°C and 110–125°C, respectively, while Thomas (1986) did not specify a growth temperature but did place the formation of blocky kaolin synchronously with the formation of late quartz overgrowths, implying a temperature in excess of 100°C. Wilkinson *et al.* (2004) noted that, comparing the diagenetic history of Giles *et al.* (1992) with the likely hydrocarbon generation times in Taylor &

Dietvorst (1991), it is apparent that the majority of the diagenesis is supposed to have occurred after peak hydrocarbon generation, i.e. after the Brent reservoirs had probably filled with oil. The temperature estimate of Giles *et al.* (1992) for the growth of the blocky kaolin is therefore considered to be too high. The blocky kaolin described by Girard *et al.* (2002) from the Hild field is said to have formed by replacement of K-feldspar, and not

by recrystallization of vermiform kaolin (which alters to illite), so that here the high temperature of formation may be genuine, perhaps influenced by evolved porewaters from the deep Viking Graben. Note that while most workers have implicitly assumed that the blocky kaolin predates the first oil charge, Wilkinson *et al.* (2004) considered the blocky kaolin to have grown during an early phase of oil charging, and Girard *et al.* (2002) placed the growth of blocky kaolin after an initial oil charge.

There are three hypotheses for the formation of the blocky kaolin. Based primarily on petrographic observations, it has been proposed that the blocky kaolin formed relatively late in the burial history, synchronous with or after the formation of quartz overgrowths. As this would be too deep for active meteoric water penetration, it has been proposed that the kaolin grew from basinal porewaters charged with CO₂ or carboxylic acids (Blackbourne, 1984; Kantorowicz, 1984; Thomas, 1986; Scotchman *et al.*, 1989). Alternatively, it has been assumed that the blocky kaolin grew from descending meteoric porewaters, either directly from the exposed surface or via a sub-surface flow path involving faults and aquifer flow (Hancock & Taylor, 1978; Sommer, 1978; Lønøy *et al.*, 1986; Glasmann *et al.*, 1989c). Thirdly, it has also been suggested that the blocky kaolin grew from stagnant meteoric porewater (a relic of the early post-depositional porewater) mixed with some acidic basinal fluid (Haszeldine *et al.*, 1992; McAulay *et al.*, 1994; Osborne *et al.*, 1994). In both the latter scenarios, it has been generally assumed that the meteoric water flow was driven by fault-related topography exposed during the early Cretaceous. It is theoretically possible that the blocky kaolin could have grown during the initial meteoric water flushing of the Brent Group during Ness or post-Ness times, but most workers have assumed that the vermiform kaolin and the calcite concretions grew during this first phase of burial, and that the blocky kaolin marks a change in conditions brought on by the early Cretaceous uplift and erosion.

Stable oxygen isotopic evidence can sometimes resolve problems of palaeo-porefluid flow. Excluding data with a significant proportion of vermiform kaolin (i.e. Osborne *et al.*, 1994 and Brint, 1989; Fig. 4), there is only a limited range of stable oxygen isotope ratios for blocky kaolin, mostly with $\delta^{18}\text{O}$ in the range of 12 to 15‰ SMOW (Fig. 4). The only dataset that is signifi-

cantly different is that of Glasmann *et al.* (1989a, their Table 2), and this may be in error. In the text of their paper, the range of kaolin $\delta^{18}\text{O}$ is given as 11.1 to 15.5‰ SMOW, but the tabulated data range from 9.3 to 11.6‰. The range stated in the text is indistinguishable from the data from the other studies, and so is probably correct. Figure 5 shows that, using the porewater evolution curve of Haszeldine *et al.* (1992), the blocky kaolin must have grown from porewaters of meteoric composition, or from a mixture of meteoric and basinal (evolved) porefluids. As quartz overgrowths do not form abundantly at temperatures below ~80°C (e.g. Fig. 3a of Bjørlykke *et al.*, 1992), then the blocky kaolin must pre-date at least the majority of the quartz overgrowths, though some growth of the two phases may have been synchronous. That kaolin grew in waters which contained at least a significant proportion of meteoric water has been

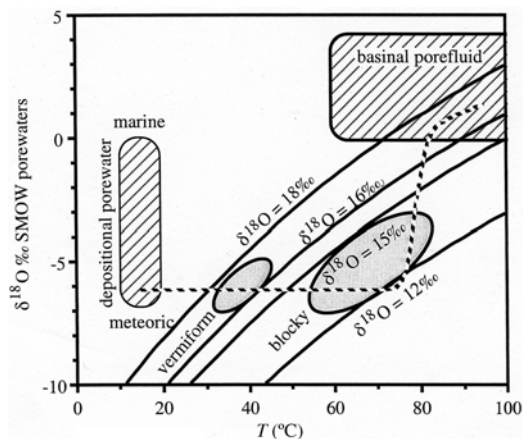


FIG. 5. Stable oxygen isotope evolution curve for Brent Group porewater (thick line), redrawn from Haszeldine *et al.* (1992). Ranges of stable oxygen isotope ratios for vermiform and blocky kaolin are taken from Fig. 4. The vermiform kaolin is constrained to have grown at low temperatures, as it is found within early calcite concretions. Therefore it is proposed to have grown in meteoric water, even in submarine fan settings such as the Magnus sandstone. The blocky kaolin, which post-dates the calcite concretions, probably grew from meteoric water or mixed basinal and meteoric water. It cannot have grown from typical evolved basinal porewater (shown on the figure) as the calculated temperatures would be too high, in many cases higher than the current reservoir temperatures which are the highest the rocks have been heated to. The kaolin-water fractionation factor is from Land & Dutton (1978).

the conclusion of the majority of the isotopic studies (e.g. Glasmann *et al.*, 1989a; McAulay *et al.*, 1994; Osborne *et al.*, 1994; Wilkinson *et al.*, 2004).

It is interesting to note that apparently identical authigenic kaolin is found in fields that are thought to have been flushed with meteoric water below the early Cretaceous unconformity, and in those that have not (e.g. Hutton Field, McAulay *et al.*, 1994; Hild Field, Girard *et al.*, 2002). Differences in the kaolin may be restricted to subtle differences in oxygen isotope composition, as in the case of the Heather and West Heather Fields. In the West Heather Field, direct exposure to lower Cretaceous meteoric water seems probable, as the Brent Group was exposed at the unconformity surface (Glasmann *et al.*, 1989c) producing kaolin with $\delta^{18}\text{O} = 15.5\text{--}16.5\%$ SMOW. The nearby Heather Field has only limited erosion at the crest, insufficient to expose the Brent Group below a covering of Heather formation shales. Here the blocky kaolin is isotopically lighter ($\delta^{18}\text{O} = 12.3\text{--}14.6\%$ SMOW; Glasmann *et al.*, 1989c), which can be explained as resulting from either precipitation at higher temperatures or from fluids with lower $\delta^{18}\text{O}$ values. As porefluids generally evolve towards heavier oxygen isotopic compositions during burial, there is no obvious source of fluids with $\delta^{18}\text{O}$ values lower than the contemporaneous meteoric water. Hence Glasmann *et al.* (1989a) proposed that the blocky kaolin grew at elevated temperatures, from meteoric water that had penetrated the sub-surface via faults, driven by a hydraulic head developed on land to the west. It is also interesting to note that the blocky kaolin of the Hild Field, which apparently formed by the alteration of K-feldspar, and not by the recrystallization of vermiform kaolin (Girard *et al.*, 2002), has stable oxygen isotopic ratios that are indistinguishable from the remainder of the northern North Sea blocky kaolin dataset (Fig. 4). It has been proposed that the Hild Field has been strongly influenced by evolved basinal porefluids derived from the graben axis (Girard *et al.*, 2002) though there is no isotopic evidence for this.

It has been commonly assumed that, if the blocky kaolin formed as a result of meteoric water penetration during the lower Cretaceous uplift and erosion, then there should be a spatial relationship between the distribution of kaolin and the unconformity surface (e.g. Harris, 1992; Bjørkum *et al.*, 1990; Giles *et al.*, 1992). While the former of

these studies claimed to have found just such a relationship, the remainder searched and failed to find any such thing. Given that the Brent Group is relatively thin (~300 m thick) and that meteoric water is thought to have penetrated to ~2 km depth (Haszeldine *et al.*, 1992), and also that total kaolin abundance is controlled by facies and detrital mica abundance (as above), than perhaps it is naïve to expect to see a simple decrease in kaolin abundance below the unconformity. It is interesting to note that Bjørkum *et al.* (1990) not only failed to find a relationship between the kaolin and the unconformity, but also that they concluded that the kaolin was of detrital origin. This has not been the opinion of later workers, or earlier of Blanche & Whitaker (1978) who considered and rejected a detrital origin for the kaolin. Interestingly, Scotchman *et al.* (1989) recorded a downward decrease in kaolin abundance within the NW Hutton field, which might have been interpreted as evidence for unconformity-related diagenesis by other authors. However, as above, the NW Hutton field was never exposed to sub-aerial erosion (McAulay *et al.*, 1994). Kantorowicz (1984) is perhaps the only worker to record a significant spatial control upon the distribution of the blocky kaolin, reporting the kaolin to be located within large sandbodies with a presumed connection to a regional aquifer.

Of which kaolin polymorph are the blocky crystals composed? Kaolinite grows at relatively low temperatures: dickite is the stable polymorph above 100°C (Beaufort *et al.*, 1998). Without specific analysis the polymorph of the kaolin cannot be determined (Hasouta *et al.*, 1999). Consequently, in at least some studies the mineralogy of the blocky crystals is not resolved. There is some confusion in the literature between kaolinite and dickite, not least over whether blocky crystals invariably equate to the dickite polymorph.

Osborne *et al.* (1994) refer to the blocky crystals as 'kaolinite', and state that XRD was used to determine the state of crystallinity, so presumably if dickite was present it would have been identified. Sommer (1978) and Kantorowicz (1984) refer to both kaolinite and dickite, but present no evidence for the presence of dickite. Durand *et al.* (2000) claim to have detected dickite within the Ness Formation of the Alwyn area. This may be based upon precise X-ray diffraction (XRD) work, or upon the assumption that blocky crystals equate to dickite. However, Ehrenberg *et al.* (1993), McAulay *et al.* (1994) and Hassouta *et al.* (1999) all

positively identify dickite in the Brent Group. It is not clear if all of the blocky kaolin crystals within the Brent Group are dickite, or if some are kaolinite. Certainly, some of the blocky kaolin crystals seem to form at much lower temperatures than the $\sim 100^\circ\text{C}$ that has been proposed for the kaolin–dickite transition (Beaufort *et al.*, 1998). Given the present uncertainty, it seems unwise to assume that all of the blocky crystals of kaolin are the dickite polymorph.

Kaolin and oil migration. Most authors have presumed that the formation of kaolin predates the first emplacement of oil into the Brent Group. Wilkinson *et al.* (2004), however, proposed that while the vermiform kaolin pre-dated the earliest oil, the transformation to blocky kaolin overlapped with an early oil-filling episode, at least in the Cormorant field. This hypothesis was based upon the spatial pattern of stable oxygen isotope data (Fig. 6). The Cormorant reservoir subsequently lost its early oil charge, allowing important diagenetic reactions to continue (e.g. the formation of quartz overgrowths) before the present-day oil charge entered the reservoir.

It has been proposed that kaolin transformed to illite during oil filling (Hancock & Taylor, 1978; Sommer, 1978). This is based on the observation that kaolin is abundant at the tops of fields, but less common at greater depths of burial where illite is

abundant, i.e. that the crestal kaolin is preserved by the oil charge. Regionally, the transformation of kaolin to illite is demonstrated by the observation that kaolin becomes less abundant over a depth range of ~ 2000 – 4000 m (e.g. Bjørlykke *et al.*, 1992; Fig. 7) while illite becomes more abundant. Kaolin is locally preserved to high temperatures where there is no K-feldspar to provide K for illite formation (e.g. 150°C in the Tarbert formation, Hild Field; Lønøy *et al.*, 1986). The formation of authigenic illite is discussed in the illite section of this paper. Relatively rapid changes in mineralogy with depth have also been observed in some fields, for example vermiform kaolinite is replaced by blocky dickite over a depth interval of only 60 m in the Alwyn Field (Hassouta *et al.*, 1999). As this depth interval is small compared to the usual interval over which this transformation takes place, it is possible that the recrystallization took place during oil filling, so that the vermiform kaolin at the top of the field was preserved by an oil charge.

Middle Jurassic – Early grain-coating illite

Several studies have observed pore-lining clay within Brent Group sandstones, that may be either detrital or authigenic (perhaps grown as smectite). Early pore-lining smectite cements have been

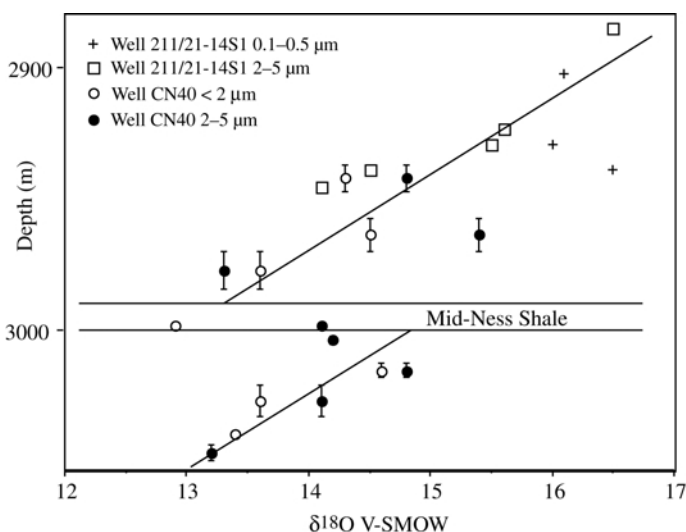


FIG. 6. Stable oxygen isotope ratio of authigenic kaolin vs. burial depth for the Cormorant field (redrawn from Wilkinson *et al.*, 2003). The linear correlation, with an offset at ~ 3000 m depth that corresponds to a low-permeability unit, the Mid-Ness shale, has been interpreted as kaolin recrystallization synchronous with early oil filling at 45 – 70°C , 80 – 50 Ma ago.

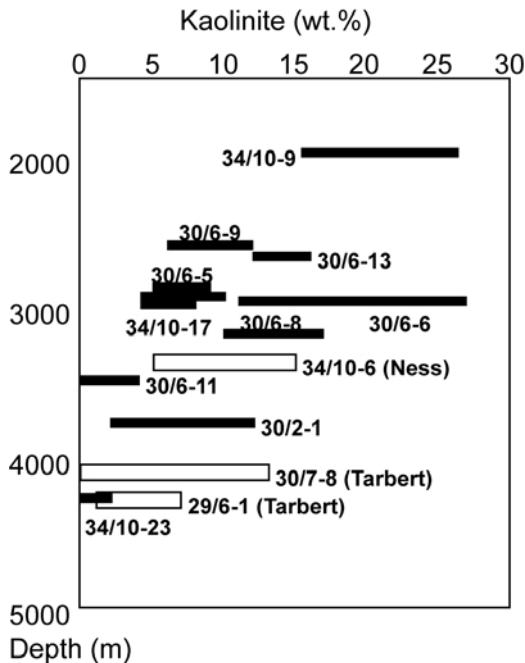


FIG. 7. Decreasing abundance of kaolin with burial depth in the Brent Group measured by XRD (redrawn from Bjørlykke *et al.*, 1992). This is due to the reaction of the kaolin with K-feldspar to form illite, which becomes more abundant over this interval.

recorded within Paleocene sands of the North Sea that contained detrital volcanic material (Watson, 1993; Watson *et al.*, 1995) and within Lower Cretaceous sands from the Scapa Field that contained siliceous biogenic debris (Hendry & Trewin, 1995). In both of these cases the SiO₂ activity in porewaters during shallow burial would have been high, significantly above quartz saturation. Kantorowicz (1984) inferred formation of early pore-lining illite within the Brent Group from interstitial saline porewaters that pre-dated the precipitation of authigenic kaolin indicative of meteoric water invasion. In this case, neither biogenic nor volcanic debris was present to elevate porewater silica levels. Kantorowicz (1984) believed that illite was stable in marine porewaters 'sufficiently saturated with silica and aluminium', a condition most likely met by the dissolution of K-feldspars or muscovite. The implication is that illite can form at temperatures as low as 15–20°C. However, grain-coating illite has also been interpreted as being formed at higher temperatures, e.g. up to 100°C (Girard *et al.*, 2002).

This interpretation may be based on the lack of pore-lining clays within calcite-cemented sandstones. It is possible that precipitation of early calcite cements is associated with the dissolution of clay minerals, along with the dissolution of feldspar grains and the etching of quartz-grain surfaces. The grain-coating illite could hence predate the calcite cement, and be early in the paragenetic sequence, and not late as inferred by Girard *et al.* (2002). There is also the possibility that the illitic coatings observed at present day were precipitated as smectite, and have subsequently recrystallized to illite or illite-smectite, see below. Such a depth-related mineralogical transformation has been documented over small depth intervals in some studies (Glasmann *et al.*, 1989a; Giles *et al.*, 1992).

Early clay coatings from Brent Group sandstones have alternatively been interpreted as detrital in origin, probably emplaced by infiltration (Giles *et al.*, 1992; McBride, 1992). In the well-sorted Etive Formation, however, a detrital origin was thought to be less convincing, unless infiltrated from the overlying Ness Formation. Giles *et al.* (1992) concluded that, in the Etive Formation, the origin of the pore-lining illite cements was 'uncertain'. Early illite-smectites were observed to be recrystallizing between 2750 and 3600 m (Giles *et al.*, 1992), developing box-work and honeycomb textures, with a platy, ragged crystal morphology. The proportion of smectite layers decreased from ~80% to 20% over the same depth interval. Comparing these observations to regional data (see below) it seems likely that the recrystallization observed by Giles *et al.* (1992) over a relatively limited depth interval is part of a progressive change in illite-smectite composition that has transformed initially quite smectitic clays to the present-day illite-rich compositions, both in sandstones and shales within the North Sea. The importance of this reaction has not been fully recognized in the literature.

Glasmann *et al.* (1989a) identified discrete smectite in shallow-buried Rannoch sandstones of the Veslefrikk Field using the XRD peak decomposition program Newmod™. They suggested that the smectite might be detrital, introduced by bioturbation, or even a contaminant introduced from the drilling mud. Given that a pore-lining clay with a box-work morphology is illustrated from the same stratigraphic interval (Glasmann *et al.*, 1989a, their Fig. 7f) it seems possible that there is an early smectite cement in this field, though this

was not identified by Glasmann *et al.* (1989a). The smectite may have been stabilized relative to illite during burial by high Ca to K ratios maintained by calcite cements within this interval, as suggested by Eberl & Hower (1976).

Several studies discuss the morphology of illite. There has been some confusion in the literature (Le Gallo *et al.*, 1998) between early anhedral platy crystals of probable detrital origin, and euhedral platy crystals developed in some highly illitized sandstones (e.g. Ellon Field, Hassouta *et al.*, 1999). In the latter case, sub-parallel arrays of platy illite are the last illite morphology to be developed during growth, forming from fibres and laths (Hassouta *et al.*, 1999). The anhedral platy clays of detrital origin act as nuclei for later, fibrous and lath crystals that grow from 1–2 μm at ~3000 m to 10–15 μm at 3350 m (Giles *et al.*, 1992).

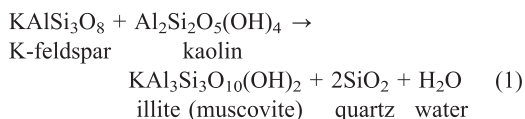
Middle-Jurassic late diagenetic illite: nucleation and precursor minerals

Late diagenetic illite can have a variety of crystal morphologies, but the most conspicuous (when viewed using an SEM on a broken chip sample) is the fibrous, or hairy, morphology. There is a fair degree of agreement between workers on the general timing of late illite growth: SEM studies show intergrown illite and quartz overgrowths, suggesting synchronous growth late in the diagenetic history (e.g. Glasmann *et al.*, 1989a, 1989c; Swarbrick, 1994; McAulay *et al.*, 1994), though some workers suggest that illite is the last diagenetic phase (McBride, 1992). Both may be true – formation of quartz overgrowth was synchronous with early illite formation. The substrate upon which the fibres grow is apparently variable. In the Heather Field, illite grows predominantly from kaolin that may be corroded but is more commonly pristine (Glasmann *et al.*, 1989c), while in other parts of the East Shetland Basin, illite grows from detrital grains or their coatings (Kantorowicz, 1990; Giles *et al.*, 1992; McBride, 1992; Le Gallo *et al.*, 1998). Replacement of kaolin by illite is also common (Kantorowicz, 1990; McBride, 1992), though vermiform kaolin is more prone to replacement than the blocky morphology (Giles *et al.*, 1992).

The optical appearance of illite in plane-polarized light (PPL) can potentially be used as a guide to the precursor mineral. Matthews *et al.* (1994) identified three classes of illite using habit and colour. Coarse

vermiform illite formed pseudomorphically after vermiform kaolin, and was colourless in PPL. Colourless groundmass illite was thought to have the same precursor phase, but to have lost the diagnostic structure. In contrast, brown groundmass illite was the result of alteration of matrix smectite, kaolin, illite and Fe oxides. It was not possible to distinguish between these illites by XRD patterns or K^+ contents; neither was there a measurable difference in the times of formation using K-Ar dating (Matthews *et al.*, 1994).

Late diagenetic illite – formation by closed-system reaction. The association between authigenic illite and kaolin has led to the suggestion that the illite forms by the reaction of K-feldspar and kaolin (Bjørkum & Gjelsvik, 1988), a closed-system reaction that may even occur within an oil-charged reservoir (Ehrenberg & Nadeau, 1989). Bjørkum & Gjelsvik (1988) presented a reaction between feldspar and kaolin which produces illite and quartz cement (equation 1).



This is a closed system reaction that does not require the import of protons (acid) or the export of any solutes, and hence does not require unrealistically large volumes of porefluid to be transported through sandstones during burial. The formula used for illite in equation 1 is actually that of muscovite, though the reaction can still proceed in a closed system for other illite compositions, with an altered ratio of K-feldspar and kaolin. The reaction above, but reversed, has been proposed by Bjørkum & Gjelsvik (1988) as a source of authigenic kaolin. The direction of the reaction (i.e. the ΔG value) is determined by the quartz supersaturation (Bjørkum & Gjelsvik, 1988), a relatively high degree of quartz supersaturation is required to drive the reaction towards K-feldspar plus kaolin. In the absence of dissolving amorphous silica, the slow kinetics of quartz precipitation below 50°C may result in high quartz supersaturations and hence the formation of authigenic K-feldspar and kaolin (Bjørkum & Gjelsvik, 1988). Above this temperature, quartz overgrowth formation is more rapid leading to lower quartz saturation in the porefluids and the formation of illite.

Evidence for the reaction of kaolin to illite comes from the Dunbar Field, where an original sub-

arkosic sand has been sampled both at shallow and deep burial depths. The deeply buried samples have almost no kaolin or K-feldspar (frequently <1%), but large volumes of illite (mostly >10% average) and quartz overgrowth. Kaolin and K-feldspar are rarely preserved within the same thin section, presumably because the reaction proceeds until one or other phase is locally exhausted. Le Gallo *et al.* (1998) modelled the reaction of kaolin and K-feldspar at ~100–110°C, and also, later, illite-producing reactions. They were able to adequately reproduce permeability and illite volume distribution in three dimensions over the Dunbar field. K-feldspar and kaolin also react together in the Middle Jurassic Garn Formation of the Norwegian sector of the Northern North Sea, producing illite (Chuhan *et al.*, 2001).

There is evidence from the Heather Field that the abundance of detrital K-feldspar is a major control on illite development (Glasmann *et al.*, 1989c), presumably because low-K-feldspar sandstones lack an internal supply of K to form illite. Reduced

detrital K-feldspar in Tarbert shoreface sandstones resulted in smaller illite contents than otherwise comparable, but more K-feldspar-rich sandstones. Since both kaolin and K-feldspar are still present within the sandstones, the illitization reaction should still be proceeding. Yet K-Ar dating has yielded illite ages from the Eocene (Glasmann *et al.*, 1989c) and different size fractions cover only a small age range suggesting only minimal detrital contamination.

Morphology. Illite morphology has an important control on the permeability of a host sandstone (Sommer, 1978; Le Gallo *et al.*, 1998; Fig 8), consequently, there are a number of studies of crystal morphology. Illite morphology can be studied *in situ* using broken chip samples on an SEM, though there is evidence that illite fabrics are damaged by standard core drying. Kantorowicz (1990) showed that critical point drying is crucial for the preservation of delicate, especially fibrous, illite fabrics. Measured permeability of core samples are higher after air drying than after

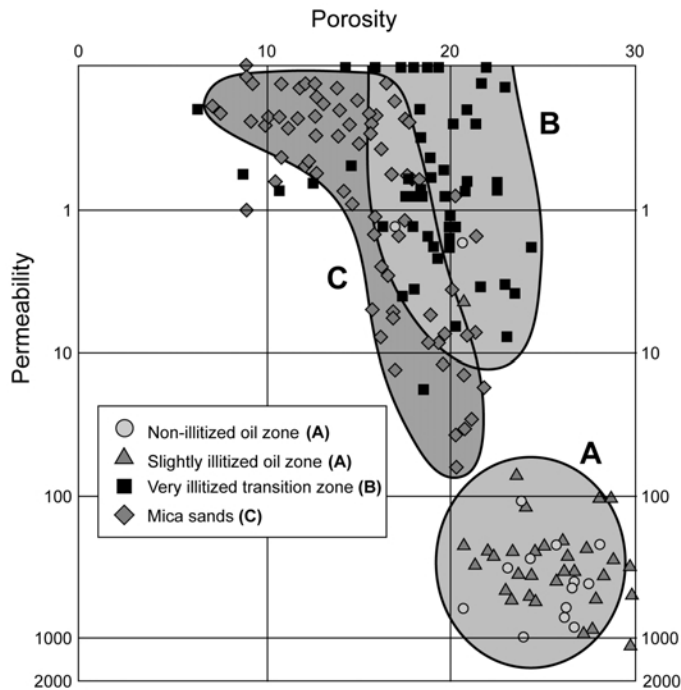


FIG. 8. The influence of illite morphology on bulk-rock permeability in the Brent Group (redrawn from Sommer 1978). Samples from Zone B have either little illite or pseudomorphic illitization of kaolin. Pore-bridging fibrous illite is not developed. In contrast, zone B has extensive fibrous illite. Zone A is 2–3 orders of magnitude more permeable than zone B. Zone C has poor permeability due to the depositional facies of the sediments, and not due to diagenetic effects.

critical point drying, as air drying tends to collapse pore-filling illite structures leaving pore networks artificially free of obstructions. This effect is greatest for illites of small particle size (Kantorowicz, 1990).

Kantorowicz (1984) suggested that illite morphology varies according to sedimentary facies, with both pore-lining and kaolin-nucleated illite in marine sandstones, but only kaolin-nucleated illite in non-marine sandstones. In contrast, there is a general theme within the literature of changing morphology of illite particles with increasing depth. This can either be over a relatively large depth range, when it is presumably related to thermally driven new growth or recrystallization, or over a small depth range when it has been related to the preservation of old morphologies due to oil charging, as suggested by Hancock & Taylor (1978). For example, Glasmann *et al.* (1989a) reported a change from fibrous to lath-like illite crystals over a present-day depth range of ~3000–4000 m in the Bergen high area.

Hassouta *et al.* (1999) examined core from the Ellon Field (Alwyn area) over a small depth range (~3200–3277 m). They recorded fibrous illite in the 'shallow' samples, laths in the intermediate ones, and plates ("plaquettes") in the 'deep' samples. There were also corresponding changes in the abundance of kaolin: there was no kaolin below 3260 m. As the temperature difference between the 'shallow' and 'deep' samples is only small, it is unlikely that the morphology changes reflect a temperature control. Given that illite K-Ar ages from Alwyn have been interpreted as recording oil charging (Jourdan *et al.*, 1987) then the morphology change could reflect the filling history. The fibres (top structure) would be oldest, the result of a short growth history terminated by early oil charging. A longer growth period, further down the structure, resulted in laths, and the longest growth formed plates.

Transmission electron microscopy (TEM) examination of highly dispersed illite separates prepared for dating usually shows two crystallite morphologies: plates and laths (Ehrenberg & Nadeau, 1989; Glasmann *et al.*, 1989b; Hamilton *et al.*, 1989; Hogg, 1989; Small *et al.*, 1992; Emery & Robinson, 1993; Fig. 9). Assuming that complete dispersion has been achieved, these are the fundamental particles of Nadeau *et al.* (1994), and should not be confused with the platy and lath-like illite

crystals observed in SEM studies, that are composed of many crystallographically-oriented fundamental particles. Emery & Robinson (1993) interpreted the two morphologies as indicating different origins: plates were detrital and laths were authigenic. They graphically estimated the end-member isotopic compositions of the two morphologies. The laths gave a reasonable authigenic result ($\delta^{18}\text{O} = 14.5 \pm 0.6\%$ SMOW), but the supposedly detrital plates had a similar composition ($\delta^{18}\text{O} = 16.5 \pm 1.2\%$ SMOW). Emery & Robinson (1993) suggested that the 'detrital' platy illite had inherited some of the ^{18}O signature from a precursor mineral, or had suffered isotopic resetting during diagenesis.

For the illite samples of Emery & Robinson (1993), there is a strong correlation between the proportion of plates and laths, and the chemistry of the separates (Al/(Al+K), Fig. 10). Lath-like fundamental particles have Al/(Al+K) = 0.68 ± 0.02 , and the plates have Al/(Al+K) = 0.46 ± 0.04 (2 standard errors). There is reasonable agreement between the composition of the laths determined from the mixing line and the ATEM results (Al/(Al+K) = 0.69 ± 0.01 ; Emery & Robinson, 1993). Do the plates have a composition which could be detrital, particularly the composition of soil-formed illite? There are not many good soil-illite analyses published as it is very difficult to obtain pure samples from clay fractions, and those published are usually corrected for a number of impurities with the consequent resulting inaccuracies (D. Bain, pers. comm.). Analyses in Newman & Brown (1987) have Al/(Al+K) of 0.59 to 0.72 ($n = 3$), and analyses of illite of weathering origin in Deer

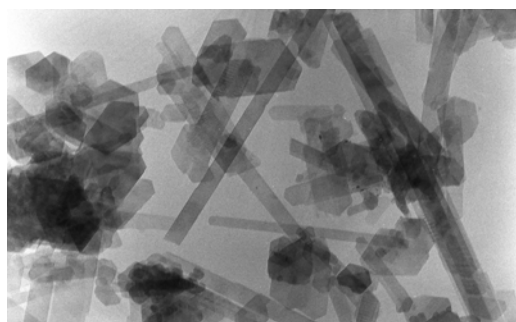


FIG. 9. TEM image of authigenic illite separated for K-Ar age dating showing typical morphology of plates and laths. Field of view = 1.8 μm . Fulmar formation, well 29/10-2, 13999 ft 9 in burial depth.

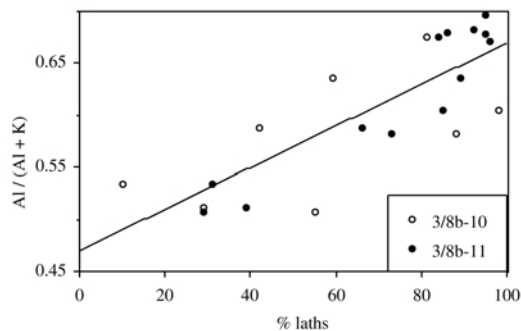


FIG. 10. Illite plates and laths have different chemistries, suggesting growth under different chemical conditions (redrawn from Emery & Robinson, 1993). The plates and laths were point counted using TEM images of illite separates prepared for K-Ar age dating. Emery & Robinson (1993) interpreted the plates as detrital and laths as authigenic in origin. Here, we interpret both crystal shapes as authigenic in origin, using both the stable isotope and geochemical data. There is some evidence that the laths pre-date the plates; see text.

et al. (1962) are similar. Hence the platy crystals lie outside the compositional range of illites of soil and weathering origin, strongly suggesting that they are of authigenic origin. In addition, Fe and Mg contents of the platy illite are also significantly different to the soil illites. While Emery & Robinson (1993) do not give full analyses, it is apparent from the very low Fe and Mg contents of illite samples containing both plates and laths that both the platy and lath-like fundamental particles must have very low Fe and Mg contents (Hogg, 1989). In contrast, the illites of soil and weathering origin have much higher Fe and Mg contents (Deer *et al.*, 1962; Newman & Brown, 1987).

It is hence probable that both the platy and lath-like fundamental particles are of authigenic origin, as suggested by the euhedral forms of the platy fundamental particles. As most studies of illite chemistry have shown increasing K with burial depth (see "Composition"), the low-K laths would be relatively early in the diagenetic sequence, with the high-K plates later. K-Ar dating of the illites confirms this (Emery & Robinson, 1993). Growth of illite under experimental conditions also produces the two particle morphologies (Small *et al.*, 1992). Lath-like illite results from high supersaturations associated with the breakdown of oxalate, with platy illite forming from lower supersaturations after the oxalate is depleted

(Small *et al.*, 1992). The role of carboxylic acids in the formation of illite within subsurface sandstones is unknown.

Timing and temperature of growth. In many areas of the North Sea, illite has been observed to be relatively rare as an authigenic phase until some 'threshold' burial depth is reached, when it becomes a major part of the clay mineral assemblage over a relatively short depth interval (Ehrenberg & Nadeau, 1989; Bjørlykke *et al.*, 1992; Giles *et al.*, 1992; McAulay *et al.*, 1993; Le Gallo *et al.*, 1998; Midtbø *et al.*, 2000; Chuhan *et al.*, 2001). Equating burial depth with temperature, it has been concluded that illite can only form at high temperatures, generally in excess of 100°C. As this is not due to the thermodynamic instability of illite at low temperatures (Bjørkum & Gjelsvik, 1988), it is attributed to the slow kinetics of illite growth. The threshold temperature for illite formation is apparently variable, from ~100–110°C in the Darwin Field (Alwyn Area, Le Gallo *et al.*, 1998) to 130–140°C in Haltenbanken (Ehrenberg & Nadeau, 1989). That a range of temperatures should be found is perhaps not surprising, given that illite formation is undoubtedly a function of porewater chemistry as well as temperature, and porewater chemistry is presumably a function of variables such as bulk rock composition. However, compilation of data from other North Sea stratigraphic units shows that illite can apparently grow over a wide range of temperatures. Even within the Brent Group, temperatures as low as 15–20°C (Kantorowicz, 1984) have been proposed for illite growth. The data which gave rise to the 'threshold temperature' hypothesis could be re-interpreted in terms of an illite-forming event, such as oil-charging or over-pressure release.

Effect of oil emplacement. There has been some debate regarding whether oil charging stops or slows diagenetic reactions. As K-Ar ages of illite are frequently close to the modelled ages of oil emplacement, many workers assume that emplacement effectively halts diagenesis (Hamilton *et al.*, 1992). However, Midtbø *et al.* (2000) concluded that illite and quartz cements had formed from residual water within the gas and oil legs within a Tarbert reservoir. The fluid inclusion data suggested that illite growth could be related to the influx of gas at 1–10 Ma ago. No K-Ar data were presented to confirm the age of the illite, which would be the youngest published from the Brent Group.

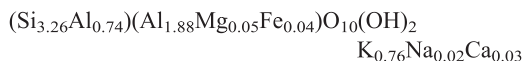
Several authors have recorded increased illite abundance in the water leg of the Brent Group compared to the oil leg (Thomas, 1986; Scotchman *et al.*, 1989; Giles *et al.*, 1992). Most have presumed that this indicates that oil charging retards or prevents the growth of illite. Contradictory evidence includes the K-Ar ages of illite from the Garn formation (Etive equivalent) which post-date modelled oil migration (using TEMISPACK; Ehrenberg & Nadeau, 1989). In addition, two cored oil-water contacts from the Garn formation show no change in illite abundance. In the Northwest Hutton field, illite abundance increases with depth over the reservoir interval (3400–3700 m; Scotchman *et al.*, 1989). This was interpreted as the result of oil charging terminating illite growth in crestral regions where the first oil charge accumulated. Illite growth in the water leg of the field ceased soon after the main migration of hydrocarbons (Scotchman *et al.*, 1989).

McBride (1992) separated clay from the sandstones of the Bruce field for isotopic analysis, most samples containing both illite and kaolin. There is no apparent change in the abundance of illite over the depth range of the study (3500–4730 m), although this is the depth range over which an increase in illite has been reported from other Brent fields. McBride (1992) observed a marked spatial control on illite abundance, suggesting that illitization was controlled by oil charging or other factors.

In many wells from the Alwyn Field the illite K-Ar ages increase with depth (Hogg *et al.*, 1987; Hogg, 1989). This has been used to infer a bottom-up filling pattern, rather than a more normal top-down one! Illite morphological trends are reversed in the Dunbar field (Alwyn area), where illite close to the structural crest is platy, and fibres are observed elsewhere (Le Gallo *et al.*, 1998).

Composition. Full chemical analyses of illite are only available for samples from relatively deep burial depths (3120–3797 m, Durand *et al.* (2000); 3550–3750 m, Hogg (1989); 3698–4539 m, Matthews *et al.* (1994); 3200–3277 m, Hassouta (1999)). Velde & Nicot (1985) and Jahren & Aagaard (1989) presented chemical data only in graphical form, on a composition plot of Velde (1977). The former paper also includes XRD determinations of smectite content. From these, it can be seen that smectite contents vary from 4–18%, and that illites with smectite contents of ~10–15% still fall within the 'illite' compositional field of Velde (1977), as do the data of Jahren &

Aagaard (1989; Fig. 11). It is probable that all of the so-called illites that have been extracted from sandstones for K-Ar age dating are actually I-S. Here, we will continue to call this mineral illite, though this almost certainly inaccurate. Hassouta (1999) gave the following average illite composition that is close to those of Durand *et al.* (2000), Hogg (1989) and Matthews *et al.* (1994):



Durand *et al.* (2000) analysed both detrital and authigenic illite, though how the two were differentiated is not stated. They found only a small difference in composition that was probably not significant compared to analytical uncertainty. Ammonium, an important interlayer cation in illite-smectites within shales (Lindgreen *et al.*, 1991), was not determined in any of the studies cited above.

There are probably too few published full chemical analyses of authigenic illite from the Brent Group to determine if there is a systematic change in composition with depth. However, K contents of illite are available for a wider depth range as these are determined during K-Ar age dating. This is useful as smectite has a minimal K content, and pure illite has K as the only cation. Hence K is perhaps the most useful element for tracking the changing composition of illite during burial.

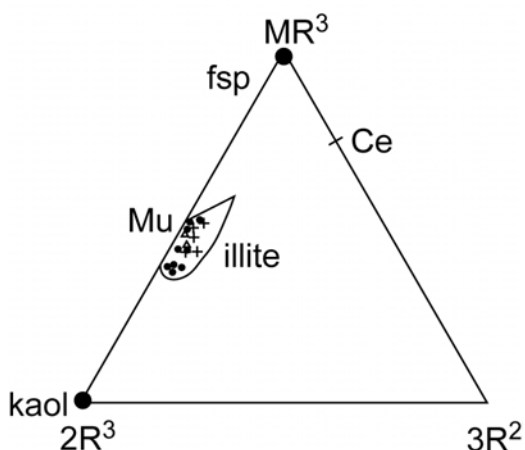


FIG. 11. Analyses of illite from the Brent Group (Jahren & Aagaard, 1989) fall in the 'illite' field on a MR^3 - 2R^3 - 3R^2 plot (redrawn from Velde, 1977). Mu is muscovite and Ce is celadonite.

Figure 12 shows that there is a systematic increase in K content of authigenic illite from the Brent Group (and other Jurassic sands) with increasing burial depth. This can be seen both in individual datasets that cover a sufficiently large depth range (e.g. Mathews *et al.*, 1994) and in the compiled data ($R = 0.61$ for 149 data points) where the correlation is significant at the 99% confidence level. Potassium contents of illite are usually measured by flame photometry, on separates of

very fine grain size (generally $<0.5 \mu\text{m}$ equivalent spherical diameter). The K content determined by this method is less than that determined on equivalent samples using an electron microprobe on polished thin sections. Clearly, the two methods analyse different samples of the illite population, which may account for the discrepancy. Girard *et al.* (2002), however, explained the difference as due to contamination of the K-Ar separates with K-free minerals, though it may be equally likely that the

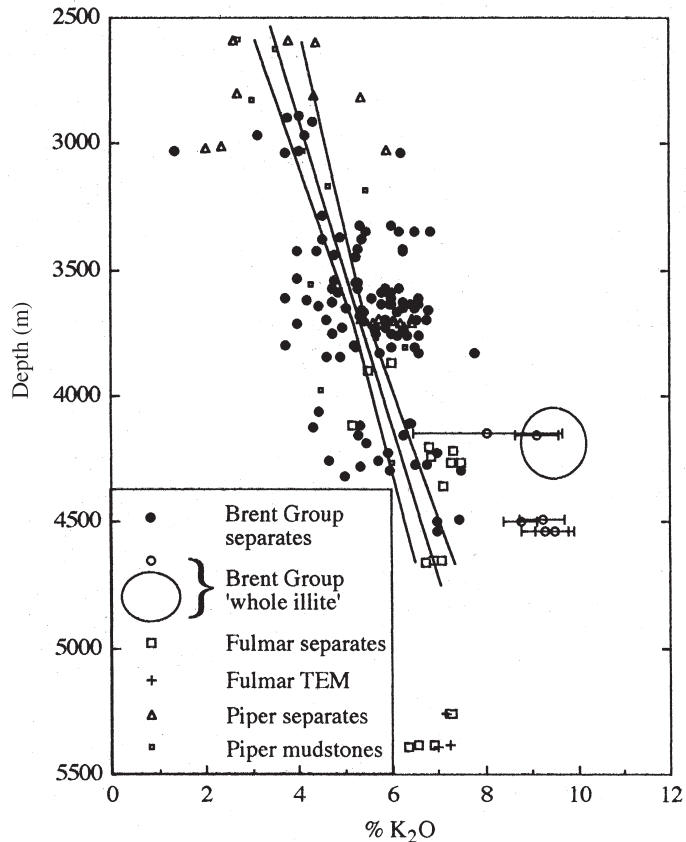


FIG. 12. Increase in K content with depth in illite prepared for K-Ar age measurement from the Brent Group (Glasmann *et al.*, 1989a; Hamilton *et al.*, 1992; Hogg *et al.*, 1993; Mathews *et al.*, 1994; Girard *et al.*, 2002), the Piper sandstone (Burley & Flisch, 1989), and the Fulmar formation (Darby *et al.*, 1994 and Table 1). All samples were illite separated for K-Ar age analysis. The best-fit line (for the depth interval 2500–4700 m) has a correlation coefficient (R) of 0.65, which for 105 points is significant for 99% confidence. The increase in K presumably reflects decreasing smectite content due either to new growth of pure illite and/or recrystallization of earlier, more smectitic illite. The data suggest that there is no significant alteration in illite composition below ~4700 m. The 'whole illite' data are *in situ* electron microprobe measurements of Mathews *et al.* (1994) and Girard *et al.* (2002), obtained from polished thin sections. The difference between the 'whole illite' electron microprobe data and the K-Ar separates may be due to contamination of the K-Ar separates, or to a genuine difference between the 'whole-illite' and the very finest size fraction separated for K-Ar analysis. The Fulmar TEM data (Table 2) were obtained from the same illite separates as the Fulmar K-Ar data. Also shown are data from mudstones from the Piper field (Burley & Flisch, 1989) which also fall on the same general trend.

'whole illite' electron microprobe analyses are sampling clay-grade authigenic K-feldspar intergrown with the illite. It is perhaps most likely that the electron microprobe analyses are unreliable as they are calibrated on porosity-free standards with a smooth, polished surface, and the illite is a high-porosity sample impregnated with resin, with a complex geometry.

The data of Hogg (1989) show a systematic change in K with grain size, with highest values recorded for the 0.5–0.1 μm size fraction. Other grain sizes (1.0–0.5 and <0.5 μm) were slightly lower, a trend often observed in clay minerals in soils (M.J. Wilson quoted in Hogg, 1989). While the lower values in the 1.0–0.5 μm fraction might be due to contamination with kaolin, the lower values in the finest grain sizes could be due to substitution of hydronium for K (Hogg, 1989), or substitution by ammonium, as above. Analysis of a single detrital muscovite sample by Matthews *et al.* (1994) found significant differences between the detrital and authigenic phases. Remembering that identification of detrital illite from authigenic illite is very difficult by the usual methods (TEM, XRD), there would seem to be some scope for assessing the degree of detrital contamination of illite separates by chemical methods (cf. Emery & Robinson, 1993).

What causes the change in the K content of the illite with depth (Fig. 12)? Given that Velde & Nicot (1985), Glasmann *et al.* (1989a), Giles *et al.* (1992) and Girard *et al.* (2002) reported significant smectite interlayers within authigenic illites, and that the former two studies reported a decrease in the proportion of smectite with depth, then either the I-S is recrystallizing to illite, or a significant volume of pure illite is growing and diluting the older I-S. Either answer seems plausible, given that the volume of illite increases significantly over the depth range of the data (e.g. Giles *et al.*, 1992) and that there is a similar change in the illite (I-S) composition within shales where recrystallization is a more plausible mechanism (Wilkinson & Haszeldine, 2002b; Fig. 12). It might be possible to use mass-balance considerations to resolve this issue, though given the considerable controversy that has arisen over the mechanism of the smectite to illite transformation, it seems unlikely that there will be a simple answer.

In general, authigenic illite is of the 1M polytype (Hogg, 1989; Jahren & Aagaard, 1989). However, with increasing burial depth, 1M illites from

Haltenbanken develop packets of the 2M polytype (Jahren & Aagaard, 1989). The latter is the thermodynamically stable polytype, suggesting that 1M illite is formed due to kinetic considerations. Given sufficient time and/or temperature, recrystallization to the 2M polytype is possible (Jahren & Aagaard, 1989).

Oxygen and hydrogen isotope data. Haszeldine *et al.* (1992) integrated illite oxygen isotope data from the Dunlin, Thistle and Murchison Fields with those from other phases to produce a well-constrained porewater evolution history for the Brent Group (Fig. 13). Illite was found to have precipitated from isotopically evolved porewaters. Data from the Bergen High area are in good agreement with the Haszeldine *et al.* (1992) data, with likely porewater $\delta^{18}\text{O}$ of +1 to +4‰ SMOW (Glasmann *et al.*, 1989a). The evolved waters were postulated to be sourced from marine water enriched in ^{18}O by diagenetic reactions within shales.

In the Heather Field, the comparison of water resistivities with oxygen isotope data suggests that isotopically heavier illite occurs in areas of higher than normal salinity (Glasmann *et al.*, 1989c, p. 277). These high-salinity waters probably have a high proportion of water expelled from deep in the basin during compaction, and are likely to have high, i.e. evolved, oxygen isotope ratios. In the Alwyn South field, Hogg *et al.* (1995) found only a small range of $\delta^{18}\text{O}$ for a large number of illite

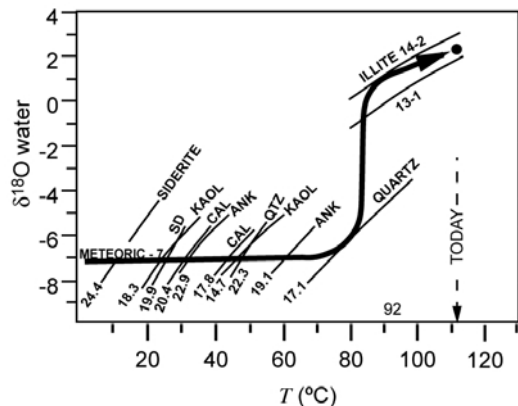


FIG. 13. Evolution of oxygen isotopic ratio in the porefluids of the Brent Group sandstones during burial, redrawn from Haszeldine *et al.* (1992). Illite was found to have precipitated from isotopically evolved porewaters. The evolved waters were postulated to be sourced from basinal water enriched in ^{18}O by diagenetic reactions within shales.

samples (+11.4 to +14.3‰ SMOW, $n = 56$) covering a range of size fractions, depth and facies. Two analyses of illite from a single shale sample gave a similar result. The time of illite precipitation was constrained to the Paleogene (Hogg *et al.*, 1995), at temperatures of 95–125°C, in porewaters with an isotopic composition close to seawater ($\delta^{18}\text{O} = -1$ to +2‰ SMOW; Hogg *et al.*, 1995) with close-to marine salinities. There is no evidence that meteoric water flushing ever occurred in this part of the Alwyn field, unlike in the examples of other Brent Group fields cited above.

Emery & Robinson (1993) measured the $\delta^{18}\text{O}$ and δD of illite from an unspecified Brent Group field. They attempted to correct the data for detrital contamination (see the section on Morphology), deducing an oxygen isotopic ratio of 14.5 ± 0.6 ‰ SMOW for the authigenic component. Combining this with K-Ar ages and basin modelling suggested illite growth at temperature of 90–130°C, and porewater compositions of -0.5 to +4.5‰ SMOW. This is again compatible with the porewater evolution proposed by Haszeldine *et al.* (1992). Emery & Robinson (1993) noted that if illite cementation was driven by fluid flow, then compactional or convective flow was implicated, and not hydrodynamic flow from the basin margins.

Interpretation of hydrogen isotope data (δD) is complicated by the possibility of resetting at relatively low temperatures. Several of the datasets show a large range of hydrogen isotope ratios compared to the range of oxygen isotope ratios (e.g. Alwyn, Hogg *et al.*, 1995; Fig. 14) that has been used as evidence for resetting. However, at least one of the Brent Group datasets shows a positive correlation between hydrogen and oxygen isotopic ratios (Bruce Field; McBride, 1992). As it is generally believed that oxygen isotope ratios are resistant to resetting at diagenetic temperatures, so the hydrogen isotope ratios must be un-reset also. It is notable that the Bruce Field data lie parallel to the trend for the Magnus Field (Fallick *et al.*, 1993; Fig. 14) interpreted to be the result of precipitation from two distinct porewaters, the origins of which are discussed in the Magnus section below. There is some suggestion that the lath-shaped fundamental particles within the illite separates have a more positive δD signature than the plates (Emery & Robinson, 1993). The cause of this is unknown.

K-Ar age dating of authigenic illite. Illite is one of the few common authigenic phases that can be radiometrically dated, due to the presence of ^{40}K

which decays to ^{40}Ar (Faure, 1986). The illite must be separated from the host sandstone, by a combination of chemical and physical techniques such as centrifuging. The aim is to obtain a very fine size fraction, in the hope that it will be free of contamination by K-bearing phases such as detrital mica. It is usually presumed that the resulting separates are the late fibrous illite observed in SEM studies of core chips.

A single age is obtained from the analysis of a single sample (in contrast to the Ar-Ar method) that must be interpreted in terms of the age of growth of the illite. There is an extensive dataset of K-Ar ages

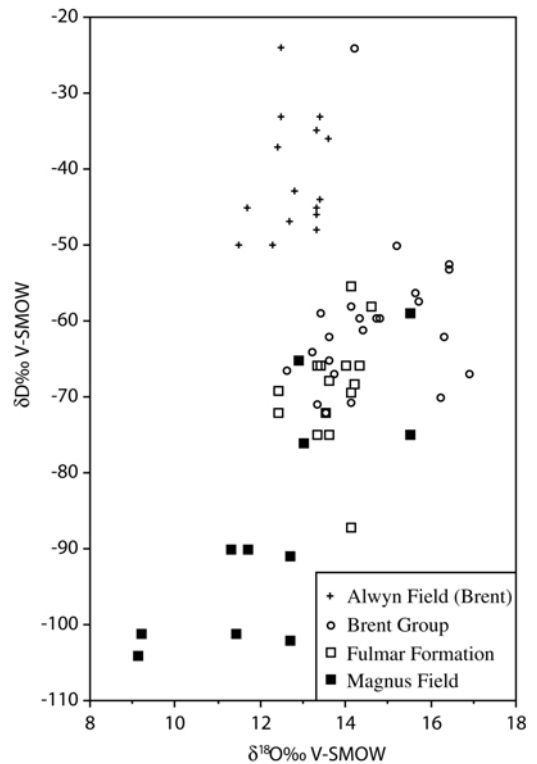


FIG. 14. Stable oxygen and hydrogen isotope ratios of illite separates from the Brent Group (McBride, 1992; Glasmann *et al.*, 1989; Emery & Robinson, 1993; Girard *et al.*, 2002) including data from the Alwyn field (Hogg *et al.*, 1995). The Magnus data (Macaulay *et al.*, 1992) and the Fulmar data (Wilkinson *et al.*, 1994) both fall on the same trend as the majority of the Brent data, suggesting that all three formations may have been influenced by porewaters of similar origin. The data from the Alwyn field are conspicuously different, suggesting a rather different fluid history.

of illites extracted from oilfield sandstones within the North Sea, and the majority are from the Brent Group. Hamilton *et al.* (1992) reviewed all then-published K-Ar ages, to which relatively little has been added since (though see Cavanagh (2002) and Girard *et al.* (2002)). Synthesizing the data into a coherent whole is complicated by the lack of coherency between the different datasets, and by under-reporting of accessory information such as burial histories, sample depths and even field names!

K-Ar ages of illite from the Brent Group sandstones range from ~60 to 17 Ma (Hamilton *et al.*, 1992; Girard *et al.*, 2002), implying that there is little or no illite growth at present day. Furthermore, K-Ar ages from a single well usually only have a small range of ages (~5–10 Ma) compared to the age of the host sandstone, even when different size fractions of illite are dated. Authigenic illite formation has hence been interpreted as an 'event' of relatively restricted duration, the nature of which is controversial. The 'event' interpretation for sandstones is in contrast with the 'thermal threshold' model which has been discussed above (see the section on 'Timing and temperature of growth'). The mechanism of growth of authigenic illite, in both sandstones and shales, is still only poorly understood (Wilkinson & Haszeldine, 2002a).

Probably the most common interpretation of illite ages is that they date oil filling of a reservoir (Hamilton *et al.*, 1992). As sample preparation for K-Ar dating involves the separation of the finest grain-size illite (separated fundamental particles in the case of the <0.1 µm grain size), is believed that the ages are of the most recently formed illite. Therefore K-Ar ages are the age of the cessation of illite growth, which in the case of an oil reservoir is the time when the water was replaced by oil. Since illite grows from the water phase by diffusion of solutes, oil emplacement effectively halts growth, or at least slows it substantially (Giles *et al.*, 1992). An example is the Heather Field, where K-Ar ages show an increase with depth (Glasmann *et al.*, 1989c). Interpreting these as the ages of filling, then the 700 m oil column took ~12 Ma to accumulate. Kantorowicz (1990) reported a change in illite morphology close to the oil-water contact in the Cormorant Field, with only grain-coating illite in the majority of the oil zone. In the water zone and the lower parts of the oil zone a further three illite morphologies were recorded (pore-bridging,

blooming and replacing kaolin), interpreted to form after oil emplacement.

While it is easy to imagine why oil filling should stop illite growth within the oil leg, it is not obvious why it should stop growth in the adjacent water leg. Yet K-Ar ages of water leg illite are commonly similar to corresponding oil leg ages (Hamilton *et al.*, 1992), despite an apparent supply of solutes and porewater saturated with respect to illite (Egeberg & Aagaard, 1989). One explanation is that the oil filling causes illite growth, perhaps by stirring the porefluids, by introducing organic acids or CO₂ (Small *et al.*, 1992; Barclay & Worden, 2000). In this hypothesis, the cessation of oil charging ends illite growth, even in water saturated areas close to the oil leg. Alternatively, illite precipitation may be triggered by the decrease of water-rock ratios at the declining stages of oil charging and associated fluid movement (Huang *et al.*, 1986).

The 'event' interpretation of illite K-Ar data has clear implications for the interpretation of petrographic data. If illite appears in a well at, say, 140°C, then this does not indicate that illite is forming at 140°C, but that it formed at some lower temperature at some point during the past (Matthews *et al.*, 1994). If both K-Ar ages, and petrographic data concerning the temperature of illitization, are available, then this can be demonstrated. One such study was conducted by Ehrenberg & Nadeau (1989) on the Garn Formation (approximate equivalent of the Etive Formation) of the Haltenbanken area. A rapid increase in the concentration of illite was observed below 3.5 to 4.0 km burial depth, corresponding to present-day temperatures of 130–140°C. Ehrenberg & Nadeau (1989) concluded that this was the temperature of illitization, and that 130–140°C was a kinetic threshold below which illite could not form at a significant rate. However, this interpretation implies that the illite should be very young. K-Ar ages from the Garn Formation ranged from 55 to 31 Ma BP, a mis-match attributed to detrital contamination by Ehrenberg & Nadeau (1989). A re-interpretation by the present authors, presuming the K-Ar ages to be reliable, suggests that illitization occurred at 110–125°C, between 31 and 55 Ma BP. Subsequent burial has not resulted in further illitization of the sandstones.

Matthew *et al.* (1994) used burial curves to estimate the depths of illitization from published K-Ar ages for the Jurassic North Sea sandstones: 1800 m (Liewig *et al.*, 1987); 2100 m (Matthew *et*

al., 1994); 2400–2800 m (Scotchman *et al.*, 1989); 2500 m (Ehrenberg & Nadeau, 1989); 3000 m (Thomas, 1986); 3000 m (Matthew *et al.*, 1994). Such a wide range of burial depths argues against a simple, thermal (kinetic) control on illitization.

The Dunbar and Alwyn North fields have an increase in illite corresponding to temperatures of 100–110°C (Le Gallo *et al.*, 1998) from K-Ar ages and fluid inclusions in quartz overgrowths interpreted to be coeval with the illite. While this could be interpreted as a kinetic barrier effect, Swarbrick (1994) has suggested, for Alwyn North at least, that illite grew during periods of hydrostatic pressuring within the reservoir sandstone, between periods of significant overpressuring. Sandstone pressure was modelled using simple 1-D models of compaction disequilibrium that were able to reproduce present day conditions. Oil migration was shown to coincide with illite growth (from K-Ar ages), with high degrees of overpressure in potential reservoirs being sufficient to exclude invading petroleum. The same apparent control of migration by reservoir overpressure was shown for the Northwest Hutton field (Swarbrick, 1994).

K-Ar ages of illites from West Heather that dated illite authigenesis at 89–94 Ma BP (Cretaceous age) were interpreted to relate to the entry of saline water into the Brent due to migration of compaction fluid along major basin boundary faults (Glasmann *et al.*, 1989c). Differences of 5–10 Ma between <0.1 µm and 0.1–0.2 µm size fractions support this interpretation. Here, illite growth is triggered by an external stimulus, such that the cessation of illite growth is caused by the withdrawal of this stimulus, and not directly by changes within the reservoir such as oil saturation or the depletion of solutes. Matthews *et al.* (1994) found that a plot of K-Ar age vs. depth for illites from the Brent Group and the upper Skagerrak Formation lay parallel to the burial curve. This was interpreted to show an illite formation ‘event’ at ~3000 m burial depth, with no subsequent recrystallization or argon loss.

One of the principal problems of dating illite is assessing the purity of mineral separates. X-ray diffraction analysis is rapid and cheap, but not reliable for low levels of contamination (<5%) that can strongly influence results (e.g. Hamilton *et al.*, 1989; Clauer & Chaudhuri, 1995). Hence some workers have used TEM examination, relying on quantitative examination of small numbers of crystals. Detrital K-feldspar is relatively easy to recognize, as the equant fragments are electron

dense, and dark on photomicrographs (Glasmann, 1992, his fig. 7a). Such clay-grade feldspar is probably the result of diagenetic corrosion of much larger detrital grains. Detrital mica can also be recognized, especially as it is frequently overgrown by authigenic illite (Glasmann, 1992, his fig. 7a). A more serious problem is the presence of detrital illite, either as separate crystallites, or as cores with authigenic overgrowths. Ideally, detrital grains should be anhedral, and authigenic grains should be euhedral, though overgrowths of detrital grains will complicate the situation. If the illite is present as discrete contaminant grains, then selective area diffraction (SAD) can be used, since detrital illite of high-temperature origin (2M) and authigenic illite (1M) are usually different polytypes (see Glasmann, 1992, his Fig. 4). Note however that Jahren & Aagaard (1989) interpreted 2M illite as authigenic in origin in Haltenbanken. A potentially more quantitative approach was attempted by Emery & Robinson (1993), using Al/K ratios, as described above.

In a review of illite ages from Jurassic sandstones of the North Sea, Clauer & Chaudhuri (1995) identified contamination of illite separates as a major barrier to the meaningful interpretation of K-Ar ages. The authors were strongly influenced by the results of Liewig *et al.* (1987), where conventional sample preparation resulted in illite separates with 5–13% K-feldspar (determined by XRD). Such high levels of contamination are unusual, but allow for graphical correction to a pure end-member illite age. Disaggregating the same samples using a commercially available freeze-dryer yielded ‘pure’ illite (by XRD), with ages that agreed well with the end-member illite ages from the contaminated samples. A more serious problem with the K-Ar method is contamination of illite separates by levels of K-feldspar too low to quantify using XRD, but which significantly affect the results (Hamilton *et al.*, 1989). An example of a dataset determined on badly contaminated separates may be the work by Girard *et al.* (2002), who presented a set of 12 K-Ar ages of fibrous illite from the Tarbert Formation. Here, TEM examination revealed detrital mica and early grain-coating illite as K-bearing contaminants. Ages ranged from 45–70 Ma, i.e. within the normal range for the Brent Group (Hamilton *et al.*, 1992). However, using petrographic relationships calibrated to fluid inclusion data, Girard *et al.* (2002) concluded that the

illite grew from 40–10 Ma, possibly coincident with oil migration at 30–35 Ma (from basin modelling), and that only the youngest K-Ar age could be used to give, at best, an estimate of the upper limit to the real timing of illite growth.

Rb-Sr age dating of authigenic illite. Compared to the K-Ar method, the Rb-Sr dating technique has only rarely been applied to authigenic illites. Liewig *et al.* (1987) attempted Rb-Sr dating of illites from Alwyn South. An isochron could not be constructed, and calculated individual ages of the samples are unreasonably high (220–350 Ma). Clauer & Chaudhuri (1995) used the data to determine $^{87}\text{Sr}/^{86}\text{Sr}$ of the porefluids from which the illite grew. Initial ratios varied from 0.7136 to 0.7239, with smaller ranges within individual wells. The data fall into two groups, one of which has a pure illite end-member ratio of 0.712, the other of which is indeterminate. Ankerite that was thought to be contemporary with illite growth had a $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.709, suggesting that at least two brines were present during illite growth.

Middle-Jurassic chlorite cements

Chlorite cements are usually only a small proportion of rock volume; however, the chlorite seems to inhibit both chemical compaction and cementation and so preserve anomalous porosity to great depths of burial (Ehrenberg, 1993). There are two types of authigenic chlorite: Fe-rich and Mg-rich (Hillier, 1994). The former are found in transitional marine to non-marine environments, such as the Middle-Jurassic sandstones of the Garn Formation in the Haltenbanken area. Here the chlorite cements form coatings upon the grains that post-date the growth of calcite concretions and mechanical compaction, but pre-date quartz overgrowths (Ehrenberg *et al.*, 1998). This is consistent with the hypothesis that chlorite does not form under earth surface conditions, but forms instead during diagenesis from an Fe-rich clay precursor such as berthierine or verdine (Worden & Morad, 2003).

The factors that control the distribution of the cements are problematic, though of obvious relevance to the petroleum industry where the prediction of high-porosity reservoirs is of economic interest. Ehrenberg (1998) used Sm-Nd age dating to test the hypothesis that the chlorite cements were controlled by sediment provenance, i.e. that chlorite was more likely to form from an

Fe-rich sediment. He concluded that there was no provenance control on the distribution of chlorite cements, and that the cements were formed by Fe-rich authigenesis on the sea floor, localized by fluvial discharge into a nearshore marine setting.

MIDDLE JURASSIC – PENTLAND FORMATION

The Pentland Formation represents a non-marine fluvial, possibly delta-top, environment (Jeremiah and Nicholson, 1999). It is an important petroleum reservoir within the Central North Sea, though in many fields it is only a secondary target after the more prolific Fulmar Formation. There are very few publications dealing with the Pentland Formation, and even fewer describing the diagenesis. Here, we draw upon our own unpublished work, which is based upon samples that are buried to between 3100 and 5800 m at temperatures of ~100–200°C.

Authigenic kaolin clay is a conspicuous feature of the Pentland formation in thin section (Fig. 15), frequently showing vermiform textures. The kaolin occurs as clusters that are approximately the same size as the detrital grains. These clusters may be pseudomorphs after feldspar, or possibly after volcanic fragments derived from the contemporaneous Rattray Volcanics. Given the non-marine depositional setting of the Pentland formation and the sub-tropical (i.e. humid) climate, it would have been reasonable to assume that the Pentland had been flushed by meteoric water at the time of deposition. The only surprising thing about the kaolin is its preservation to high temperatures, given that illite is generally more stable than kaolin above ~100°C, as in the Brent Group, above. In

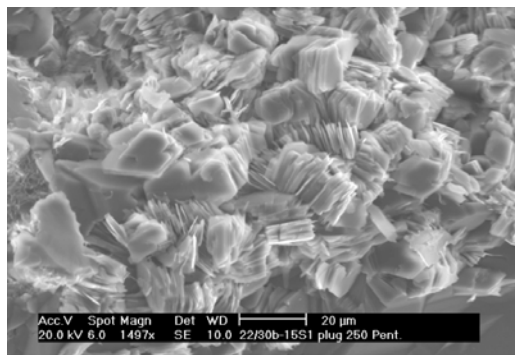


FIG. 15. Photomicrograph of kaolin from the Pentland formation.

order to convert kaolin to illite, there must be a supply of K, as illite contains significant K but smectite has none. Conventionally, K is supplied by the dissolution of K-feldspar, which is generally lacking in the Pentland Formation (0–5%, unpublished data). The Pentland Formation is therefore deficient in K relative to a ‘normal’ sand. This would necessarily restrict the reaction of kaolin to illite, stabilizing kaolin at high temperatures at which it would normally not survive. There has been no detailed mineralogical study of the Pentland Formation published, consequently there is no data concerning the possible presence of dickite.

UPPER JURASSIC – CENTRAL NORTH SEA (FULMAR FORMATION)

The Fulmar Formation is the principal reservoir of the Central North Sea. Surprisingly little work has been published describing the diagenesis, despite the sometimes extreme burial depths (down to 7 km) that mean that diagenetic modifications to reservoir properties can be extreme, even to the extent that most hydrocarbon production is from secondary porosity (Haszeldine *et al.*, 1999). In the section below, we draw upon our own unpublished data. The diagenetic sequence of the Fulmar field, in which the reservoir is the Fulmar Formation, was described by Stewart (1986). As this field is buried to only ~3000 m, diagenetic clays were not seen as volumetrically important, though this is not true in many of the deeper sections.

Kaolin clay is largely absent from the Fulmar Formation (Stewart, 1986; Wilkinson *et al.*, 1994). Was kaolin formed during shallow burial and subsequently altered to illite clay, or was it never formed? Stewart (1986) presumed that kaolin had never grown, and used this as evidence that porewater chemistry had never favoured kaolin formation. However, small quantities of kaolin are occasionally observed within the Fulmar Formation, and XRD analysis of the clay fraction commonly detects significant quantities of this mineral (unpublished data). The issue is therefore not straightforward, and is of interest for two reasons. Firstly, kaolin can react with K-feldspar during deep burial, to form illite, leaving potentially important secondary porosity. Also, kaolin is often regarded as forming during early burial in meteoric water, and so is evidence of meteoric water ingress

into a sandstone (Fairbridge, 1967). For a deeply buried marine sandstone (such as the Fulmar Formation), it can be difficult to assess if early meteoric water flushing has occurred (Wilkinson *et al.*, 1994). As feldspar dissolves during meteoric flushing, forming secondary porosity, it is important to know if this process has occurred as part of understanding the porosity evolution of a sandstone. There are several lines of evidence to consider:

(1) There are traces of kaolin in the Fulmar even at extreme burial depths, detected by both XRD and optical methods (unpublished data). This could be the remnant of a more abundant, shallow cement phase.

(2) As the Fulmar sea was bordered by land, from which ran rivers supplying the sediment (Jeremiah & Nicholson, 1999), then rainfall on the land would have entered the groundwater, and ‘meteoric lenses’ of water would have penetrated into the shallow buried Fulmar sediment.

(3) Kaolin is found in the more shallowly buried Fulmar on the Western Platform, where burial depths are not so extreme (<3000 m; unpublished data).

(4) If an attempt is made to reconstruct the oxygen isotopic evolution of the porewater within the Fulmar during burial, then meteoric water flushing is required, at least for well 29/10–2 (Fig. 16). This well is currently at ~4400 m burial depth, i.e. intermediate in the regional context.

It has been suggested that the deepest areas of the Central Graben may never have been subjected to meteoric water flushing, presumably because they were too remote from a palaeo-shoreline to be invaded by a ‘meteoric-lens’ (e.g. Elgin and Franklin Fields; Hendry *et al.*, 2000). This hypothesis was based on the preservation of detrital plagioclase that has been described as ‘pristine’. As plagioclase tends to dissolve in meteoric water, this is evidence that meteoric flushing (and kaolin formation) never occurred. However, on balance, evidence suggests that the majority of the Central Graben was once flushed by meteoric water, probably as part of a ‘meteoric lens’ attached to the western shoreline. Kaolin would have been abundant, but has subsequently altered to illite, leaving only remnants. It is possible that the deepest parts of the graben were not flushed by meteoric water, or perhaps not for sufficiently long to form significant kaolin at the expense of plagioclase or K-feldspar.

The clay mineralogy of the Fulmar Formation is dominated, at present day, by illite. This includes

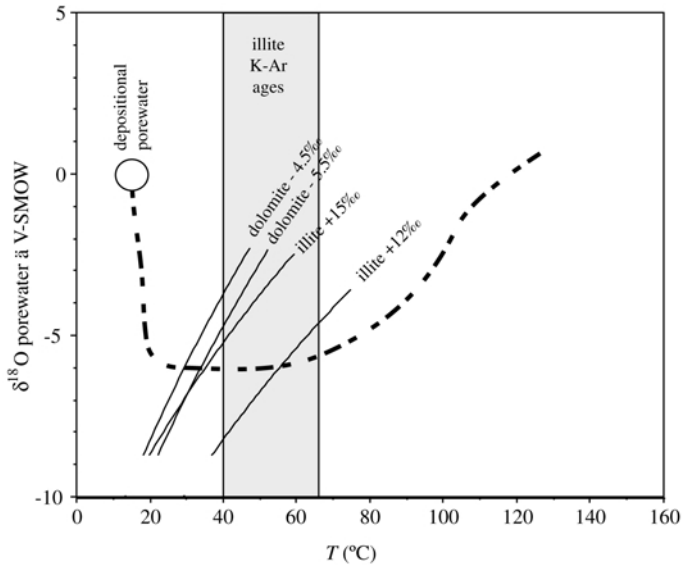


FIG. 16. Porefluid evolution diagram from the Fulmar formation in well 29/10-2. While the early part of the curve is constrained by the stable oxygen isotopic data for very early dolomite and illite (illustrated), the later part of the curve is drawn using unpublished data from other Fulmar fields. Note the very low temperatures of illite formation, derived from K-Ar age dates and the regional geothermal gradient. Fractionation equations are from Land (1983) for dolomite and Yeh & Savin (1977) for illite.

both authigenic, neoformed fibrous illite, and grain-rimming clay. Some of the latter is probably recrystallized detrital clay, some of which may have originally been faecal pellets, now dispersed by the ubiquitous bioturbation. The original mineralogy of the detrital clays is unknown, though was probably smectite or illite-smectite, as described for the Brent Group, see above. The authigenic fibrous illite has been described by Wilkinson *et al.* (1994) from well 29/10-2, that lies some 7 km south of the Puffin Field. Late fibrous illite is an important authigenic phase, intergrowing with quartz overgrowth (Wilkinson *et al.*, 1994). The illite is generally nucleated upon the detrital clay coatings of framework grains, except

where it is replacing authigenic kaolin. A late bitumen 'cement' covers much of the illite, and makes study difficult.

K-Ar ages of authigenic illite (presumed to be the late fibrous morphology) fall into two groups (Darby *et al.*, 1997 and Table 1; Fig. 17). The earlier group was interpreted as the product of compactional porewater movements due to a sudden increase in the subsidence rate, while the younger were linked to either periods of overpressure release or to oil migration (Darby *et al.*, 1997). As the K-Ar ages within well 29/10-2 become younger with depth (Fig. 18), illite growth is now proposed to have occurred during slow filling of the structure with hydrocarbons in the late Cretaceous, between

TABLE 1. K-Ar age data for illite separates from the Fulmar Formation, well 29/5b-6.

Depth (m)	Grain size (μm)	Radiogenic ^{40}Ar (10^{-10} mol/g)	Radiogenic % ^{40}Ar	K (wt.%)	Age \pm σ (Ma)
5278	<0.1	4.364	24.49	7.29	34.2 \pm 1.7
5400	0.1–0.05	6.571	63.33	6.91	54.0 \pm 1.3
5405	0.1–0.05	3.785	19.33	6.56	33.0 \pm 2.1
5410	0.1–0.05	5.969	43.95	6.36	53.3 \pm 1.6

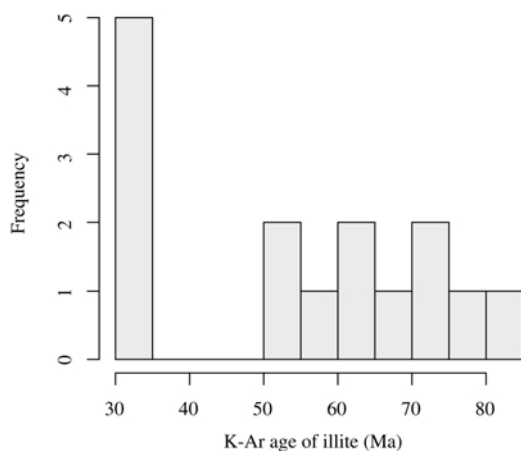


FIG. 17. Histogram of K-Ar ages of authigenic illite from the Fulmar Formation (data from Darby *et al.*, 1997 and Table 1).

84 and 58 Ma. The well is currently 'dry' i.e. does not contain hydrocarbon in commercial quantities, so that the original hydrocarbon charge must have escaped subsequent to illite growth. Residual bitumen proves that hydrocarbon has been present in the well. K-Ar ages (Darby *et al.*, 1997 and Table 1) imply illite growth at temperatures of only 40–65°C in porewaters of meteoric origin (Fig. 16). This fits the majority of the stable oxygen isotopic data. This temperature of illite growth is low compared to the majority of values reported for the Brent Group (see above), though within the range of reported temperatures (15–140°C).

The composition of the illite within the Fulmar Formation plots on the same % K-depth trend as the Middle Jurassic data (Fig. 12). This suggests that here also, the 'illite' is actually I-S (though there is only a minor swelling component at such depths), and that the percentage of smectite decreases during burial either due to recrystallization or new growth unaccompanied by dissolution. The increase in K is

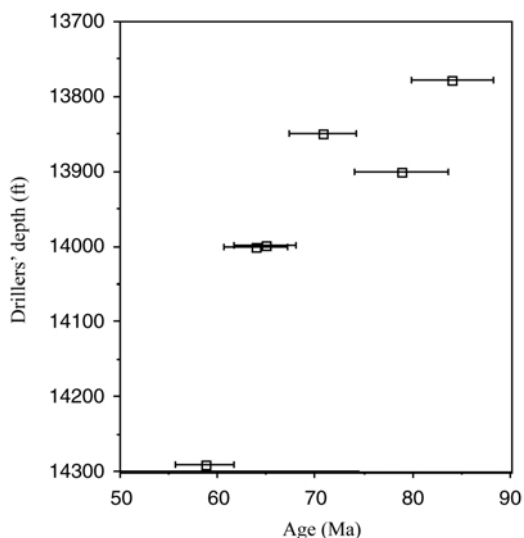


FIG. 18. K-Ar age vs. burial depth for authigenic illite from the Fulmar formation in well 29/10-2 (data from Darby *et al.*, 1997). The systematic increase in depth is taken as evidence that the illite grew during oil filling of this reservoir. Estimated temperatures at the time of filling are low, only 40–65°C.

not apparent in the data from below ~4500 m, suggesting that the recrystallization/new growth reaction may have stopped. There is good agreement between K contents derived by flame photometry for K-Ar age dating and by analytical TEM (Table 2, Fig. 12). There is a small discrepancy between K contents of the illite separates determined by flame photometry and atomic absorption spectrometry (Table 1, Fig. 6) though not sufficient to alter the interpretation of the data.

Stable oxygen and hydrogen isotopes of illite extracted for K-Ar age determination show very similar values to those from the Brent Group (Wilkinson *et al.*, 1994; Fig. 14). Although Wilkinson *et al.* (1994) interpreted the hydrogen

TABLE 2. ATEM analyses of illite prepared for K-Ar age dating from the Fulmar Formation. Results as element wt.%. Each is an average of four to six analyses.

Depth (m)	Grain size (µm)	K	Mg	Al	Si	Ca	Ti	Fe
5278	<0.1	7.2	1.1	13.1	15.5	0.5	0.46	2.6
5400	0.1–0.05	7.3	1.1	13.3	15.8	0.6	.55	2.2
5405	0.1–0.05	7.3	1.1	13.1	15.5	0.5	0.49	2.8
5410	0.1–0.05	7.0	1.1	12.8	14.8	0.4	0.4	4.0

isotopic ratios to be reset during burial (on grounds of the large spread of hydrogen data when compared to the oxygen data) it is apparent that instead, the two-fluids model proposed for the Upper Jurassic Magnus fan sandstone (Macaulay *et al.*, 1992) could equally be applied, as described below.

UPPER JURASSIC – SUBMARINE FAN SANDSTONES

Piper Formation (Outer Moray Firth)

The Piper Formation of Oxfordian–Kimmeridgian age underlies the Kimmeridge Clay source rocks in the Outer Moray Firth (Fig. 1). Burley & Flisch (1989) proposed a complex diagenetic history for the Piper Formation sandstones involving multiple phases of illite growth. An early burial (eogenetic) phase of pore lining illite or I-S cementation identified by Burley (1986) and Spark & Trewin (1986) was preserved within the oil zone and dated as 151–93 Ma using the K-Ar dating method. Water zone dates were dominated by late stage fibrous illite grown during, or shortly after, oil emplacement (mostly 40–70 Ma). These ages required correction for detrital K-feldspar contamination, yielding corrected ages of 30–65 Ma (Burley & Flisch, 1989). Oil accumulation was considered to have taken place ~29–35 Ma ago. Potassium contents of the Piper illites, as extracted for K-Ar age determination, fall on the regional trend of increasing K with depth (Fig. 12). This implies a significant smectite component in the ‘illite’, and that the proportion of smectite decreases with depth due to growth of new, pure illite, or by recrystallization of existing smectite to illite. The Piper mudrocks are discussed below.

Piper Formation (Inner Moray Firth)

Within the Inner Moray Firth, the sandstones of the paralic Piper Formation have both vermiform and blocky kaolin (Spark & Trewin, 1986) but only minor illite. No attempt was made to elucidate the relationship between the two kaolin morphologies. In contrast, kaolin is present only rarely in the turbiditic Ten Foot sandstone enclosed within the Kimmeridge Clay Formation and the Claymore Sandstone member turbidites. This can be attributed to the lack of meteoric water flushing in turbidite units. Diagenesis was controlled by the chemistry of

the depositional porewaters (Spark & Trewin, 1986). Kaolin is also present within the mudrocks of the Piper formation (Burley, 1986) though whether it is authigenic or detrital was not determined.

An arenaceous sequence of Oxfordian age from the Inner Moray Firth may hold the only quantitatively abundant early illite cement recorded within the UK offshore area. Although absent from marine sandstones, the cement constitutes up to 20% of the solid rock volume in non-marine facies (Jeans & Atherton, 1989). Described as a ‘mica’, the 10 Å clay occurs as thin fibrous or platy crystals lining many primary and secondary pores. K-Ar dating suggests growth from 97–115 Ma, concurrent with the argillization of the eroded granite upon which the sandstones were deposited. Associated mudstones may also contain authigenic illite (Jeans & Atherton, 1989).

The illite cements must have formed at unusually low temperatures. The K-Ar ages correspond to burial depths of only ~1200 m, implying temperatures of 35–40°C. Organic maturity data from surrounding mudrocks precludes illite growth during a ‘thermal event’. Jeans & Atherton (1989) suggest that some of the sandstones may have included unstable volcanic detritus, and that this may have caused the formation of early siliceous and illitic cements, in a closed diagenetic system.

Magnus Sandstone

The Upper Jurassic Magnus sandstone was deposited on a submarine fan fed from the uplifted Scottish landmass. The kaolin within the Magnus sandstone is similar to that recorded in other North Sea formations such as the Brent Group. Both vermiform and blocky kaolin are recorded, with the vermiform morphology most common at the crest of the field, and blocky kaolin more common on the flanks (Macaulay *et al.*, 1992). Such a trend could be re-interpreted as kaolin recrystallization during progressive oil filling of the reservoir, as in Wilkinson *et al.* (2004). There is sufficient dissolved feldspar within the Magnus sandstone to account for the formation of all the observed kaolin in a closed system (Macaulay *et al.*, 1992).

The Magnus Sandstone has the only (blocky) kaolin isotopic data that are significantly different from the other North Sea datasets (Fig. 19; Macaulay *et al.*, 1992). The stable oxygen isotopic data are indistinguishable from the rest of the

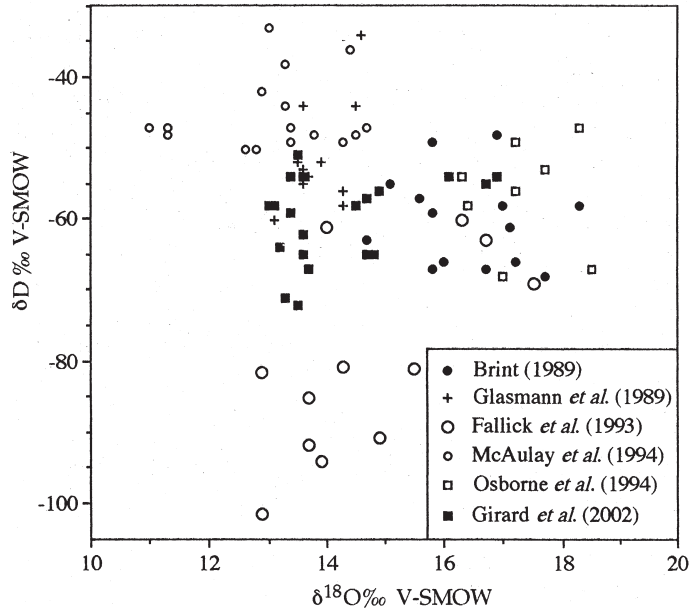


FIG. 19. Cross plot of stable oxygen and hydrogen isotope ratios for kaolin from the Brent Group and Magnus sandstones. The range of oxygen ratios is explained by some of the samples being vermiform kaolin (Osborne *et al.*, 1994) or mixtures of vermiform and blocky kaolin (Brint, 1989). The remainder of the data are from blocky kaolin, and show a remarkably small range of oxygen isotope ratios. For the hydrogen isotope ratios, the range is again fairly small, with the exception of the Magnus sandstone data (Fallick *et al.*, 1993). This could indicate resetting of the hydrogen ratios, but has instead been interpreted as the influence of a porewater with very light hydrogen isotopic ratios, of unknown origin.

datasets, but the δD data have a unique range (Fig. 19), with extreme values less than -100‰ SMOW (Macaulay *et al.*, 1992). As $\delta^{18}\text{O}$ and δD correlate, the isotopic data can be explained as a mixing line between two end-member porefluids. However, while one fluid is plausibly a typical evolved basinal fluid, the other would be of unusual and unexplained composition. It is possible that the hydrogen isotopes are reset by (for example) interaction with organic material, though this would require resetting of oxygen isotopes which seems unlikely (Fallick *et al.*, 1993). Macaulay *et al.* (1992) interpreted the isotopic data to indicate the growth of blocky kaolin and illite from stratified waters over a period of ~ 35 Ma.

There are two phases of illite precipitation recorded from the Magnus Sands. Both Rainey (1987) and Macaulay (1990) observed early, grain-coating illite and a later fibrous illite contemporaneous with quartz overgrowth formation. The early illite consists of 'hand-shaped' flakes perpendicular to the grain surfaces with wispy ends with anhedral terminations that are likened to the fingers of a

hand. Rainey (1987) interpreted the anhedral ends as an artifact of air-drying of the samples, while Macaulay (1990) suggested that they represented dissolution. The identification of the clay as illite (as distinct from smectite) was based upon EDS measurements of the clay composition, and especially the presence or absence of K. With the very small particle size, accurate determination would be impossible. Hence it is possible that the clay was smectite, though it may now be partly illitized.

The late-stage illite grew contemporaneously with quartz overgrowths, and was nucleated either upon the early illite or grew by the progressive illitization of orthoclase and kaolin (Rainey, 1987; Macaulay, 1990). Some of the most intensely illitized kaolin and feldspars occur within the oil zone, while the least altered kaolin is found within the water zone (Rainey, 1987). This was interpreted to suggest that illitization was complete before oil charging (Rainey, 1987). In contradiction, De'Ath & Schuyleman (1981) and Heavyside *et al.* (1983) both recorded more illite in the water leg of the reservoir than the oil leg, interpreted as indicating

that oil emplacement had halted diagenesis. McHardy *et al.* (1982) also interpreted illite growth to be terminated by oil emplacement, and Macaulay (1990) observed more illite in down-dip wells than in crestal ones. K-Ar ages (see below) suggest that illite continued to grow in the water-leg after oil emplacement (Macaulay, 1990).

Fundamental particles of illite separated from Magnus sandstones for K-Ar dating are thin (20–30 Å; McHardy *et al.*, 1982; Nadeau, 1985), i.e. 2–3 unit cells thick. Kantorowicz (1990) interpreted this as due to the relatively short period of time the Magnus reservoir has spent above 100°C, only 10–25 Ma. The precipitation of illite within the Magnus sand may have been triggered by the influx of CO₂ associated with the early stages of oil generation. Using geochemical modelling, the influx of CO₂ was shown to cause the dissolution of K-feldspar and the precipitation of quartz, ankerite and illite within an arkosic sand host (Barclay & Worden, 2000).

Oxygen isotopes. Porewaters within Magnus Field apparently remained stratified and stable over at least 35 Ma, during upper Cretaceous and Paleogene times (Macaulay *et al.*, 1992). Separated minerals (illite and kaolin, also siderite and ankerite) show a consistent difference of ~3‰ in δ¹⁸O between crestal and flank wells. In addition, isotopic data (δ¹⁸O and δD) for both illite and kaolin show a strong linear correlation (Fallick *et al.*, 1993; Fig. 19), interpreted as precipitation of minerals from fluids of variable composition at constant temperature (Fallick *et al.*, 1993). This implies the presence of two isotopically distinct porefluids within the Brent sandstone, one of which might be the present-day Magnus reservoir fluid, but the other is more problematic. Fallick *et al.* (1993) suggested that a Cretaceous meteoric porewater with altered δD values could be implicated. The meteoric water is present predominantly within the crestal area of the field, consistent with entry during Cretaceous uplift and erosion. The diagenetic record ends ~55 Ma ago, when K-Ar ages of illite in the water leg record the cessation of illite growth.

K-Ar age dating of illite. Authigenic illite from the Magnus Sandstone yields ages of 88 to 42 Ma (Macaulay *et al.*, 1992; Emery *et al.*, 1993). The illite in the oil leg (youngest age 55 Ma) was thought to record oil filling (Macaulay *et al.*, 1992) while a younger age in the water leg reflected continued illite growth after oil filling. The oldest

recorded age (88 Ma) was rejected as probably due to sample contamination (Emery *et al.*, 1993). However, a recent re-examination of this field, along with basin modelling, has shown that this age is coincident with oil migration into the field (Cavanagh, 2002). There are now thought to be two kitchen areas providing oil to the Magnus structure, which resulted in a spread of illite ages. Cavanagh (2002) related all the K-Ar ages to oil-filling events.

The Brae area

Authigenic kaolin is present within both sandstones and mudrocks of the open marine Heather Formation of the Brae area (Greenwood *et al.*, 1994). The earliest kaolin is associated with altering detrital micas, with vermiform kaolin apparently slightly later in the paragenetic sequence. Figure 3C of Greenwood *et al.* (1994) illustrates kaolin that is very similar to the blocky kaolin described from the Brent Group (see above), though Greenwood *et al.* (1994) do not distinguish this from vermiform kaolin. It is possible that the process of kaolin recrystallization described from the Brent Group also occurred within the Heather Formation. The kaolin is described as being an alteration product of detrital feldspar. There are no data concerning the presence or absence of dickite.

Kaolin is not present within the pro-delta sands of the Hugin Formation (Greenwood *et al.*, 1994), even though it is present within the laterally equivalent Heather Formation, as above. The difficulty is not, perhaps, so much to explain the absence of kaolin in the Hugin Formation (which is unstable with respect to illite at the depths the study material was taken from) but to explain the presence of kaolin in the Heather Formation. Presumably the Heather Formation is deficient in K, without which the illitization reaction cannot proceed, as with the Pentland Formation of the Central North Sea, discussed above. The distribution of kaolin is not simply controlled by the distribution of meteoric porewaters during early burial, otherwise kaolin should be present within the marginal marine Hugin Formation and absent from the fully marine Heather Formation. This is the reverse of the observed pattern.

In the Brae oilfield, the reservoir sandstones are submarine fans enclosed within the argillaceous Kimmeridge Clay formation. The paragenetic sequence for the sandstones of the South Brae

oilfield proposed by McLaughlin *et al.* (1994) is unusual in that the kaolin is proposed to be late, apparently post-dating the majority of quartz overgrowth formation. Also unusual is the lack of clay minerals present compared to the volume of K-feldspar that has apparently dissolved – there is so little authigenic clay that it was proposed that both Al and Si had been exported from the Brae formation (McLaughlin *et al.*, 1994). It was suggested that early meteoric water flushing removed Si and Al so completely from the formation that no early kaolin was formed, in contrast with all the other formations in the area. Later, basal waters that were forced through the aquifer during compaction of the basin were proposed to have been charged with organic acids that mobilized Si and Al from feldspar dissolution and removed them from the sandstone. Hence, in this case, kaolin is proposed to have been largely absent from the formation because it never formed.

In another shallow-marine sandstone enclosed entirely by the Kimmeridge Clay formation, Jeans & Fisher (1986) also found kaolin to be the last authigenic silicate phase. However, at least some of the kaolin is enclosed by a concretionary calcite cement, which presumably post-dates the kaolin. Again, kaolin is relatively uncommon. The dissolution of feldspars to form kaolin was related to the biotic generation of organic acids and the abiotic generation of CO₂ (Jeans & Fisher, 1986). In another similarity to the study of the Brae field sandstones (McLaughlin *et al.*, 1994; above), Jeans & Fisher (1986) record two phases of silicate dissolution with no apparent reaction products.

The sandstones of the Brae area have two generations of illite. The earlier phase is grain-coating, and the later, fibrous and porefilling (McLaughlin *et al.*, 1994; Greenwood *et al.*, 1994). McLaughlin *et al.* (1994) described the Kimmeridgian–Volgian Brae Sandstone, while Greenwood *et al.* (1994) worked upon the Bathonian to Oxfordian Hugin, Sleipner (Pentland) and Heather Formations. The paragenetic sequences presented by the two teams differ significantly in the order of growth of authigenic phases, although both published studies place the pore-lining illite after significant diagenetic events (calcite concretion and early quartz overgrowth formation). The illite would hence be of authigenic origin, rather than being infiltrated detrital material. However, as noted elsewhere, concretionary calcite can cause the dissolution of earlier phases resulting in spurious

petrographic relationships, so that there is at least some uncertainty regarding the origin of the pore-lining illite. The paragenetic sequence of McLaughlin *et al.* (1994) shows that the illite clay rim cement developed in mid-diagenesis, after the growth of calcite concretions and contemporaneous with the onset of quartz overgrowth formation. This implies temperatures of 50–90°C (McLaughlin, 1992). The fibrous, late illite is only a minor phase in the Brae Formation (McLaughlin *et al.*, 1994). A single sandstone K-Ar age of 48.5 Ma from the Sleipner Formation was deemed to be typical of the North Sea Jurassic sandstones, and to represent a growth temperature of ~100°C (Greenwood *et al.*, 1994). Stable oxygen isotope data of illites (two analyses) suggested precipitation from evolved porewaters of marine origin (+2‰ SMOW; Greenwood *et al.*, 1994).

CRETACEOUS

There is remarkably little published work concerning the clay minerals of the thick Cretaceous section from the North Sea, partly due to the fact that much of the rock is chalk! However, siliciclastic rocks do exist, and Lower Cretaceous shales form commercially important seals to Jurassic reservoirs in many places. There is also a small siliciclastic component in chalk, and this has been studied in detail, as below. Maliva *et al.* (1999) described authigenic clay minerals from within veins in the chalk, and suggested that they had been sourced from lower down in the stratigraphy.

Hendry & Trewin (1995) described a lower Cretaceous shale, ~100 m thick, from the Scapa Field of the Witch Ground Graben (Fig. 1). At burial depths of ~2500 m, the mineralogy was described as being dominated by smectite and illite, with subordinate chlorite and kaolin (Jones *et al.*, 1989). Presumably the smectite is actually present as I-S, with or without additional discrete illite. The method of identifying clay phases from the XRD traces was not described.

The clastic component of chalk can only be studied after extraction from the carbonate matrix by acid dissolution or where concentrated as clay layers, that are thought to be at least partly diagenetic in origin (Lindgreen *et al.*, 2002). Around 90% of the clay is a high-smectite I-S phase and a low-smectite illite-smectite-chlorite (I-S-Ch) phase, probably of detrital origin.

However, two samples contain significant amounts of tosudite and Ch-Serpentine. Diagenetic reactions involve fixation of Mg (possibly sourced from high-Mg calcite) and neoformation of a tosudite or of a random mixed-layer trioctahedral chlorite-berthierine.

MUDROCKS

The mudrocks of the North Sea are the principal oil-source rocks, occurring in greatest thicknesses in the Viking and Central Grabens. Interlayered illite-smectite (I-S) is an important component of mudrocks of Jurassic to Paleogene age. The I-S has been interpreted to be at least partly of authigenic origin, the product of the alteration of unstable basic volcanic debris (Pearson *et al.*, 1983). In detail, the I-S has a vermiculite component (see below). While there are only limited chemical data from the North Sea, there is a systematic increase in K in the I-S in the shales of the Piper field (Burley & Flisch, 1989; Fig. 12). This is indistinguishable from the compositional change in the illite within sandstones over the same interval, and presumably reflects an ongoing smectite-to-illite transition, as below.

Mudrocks throughout the Jurassic sequence have only a small percentage of smectite layers within illite-smectite (~20%, Wensaas *et al.*, 1994). However, there is reasonable evidence that the I-S was originally more smectitic and has been illitized during diagenesis. Several studies have documented a progressive increase in the illite content of I-S with depth through North Sea Paleogene–Mesozoic sections, from a wide geographic area including the Viking Graben–Moray Firth (Pearson *et al.*, 1982; Pearson *et al.*, 1983; Pearson & Small, 1988; Burley & Flisch, 1989; Wensaas *et al.*, 1994) and the Central Graben (Dypvik, 1983; Lindgreen *et al.*, 1991). For example, within the mudrocks overlying the Brent Group in the Gulfaks field there is a clear trend from highly smectitic clays at shallow burial depth (70–100% smectite) to more illitic clays at depths of 1.5 to 3.5 km (Wensaas *et al.*, 1994). This trend is similar to the change in mudrock composition in the Texas Gulf coast, which has been the source of much controversy since the work of Hower *et al.* (1976). The Brent Group mudrocks also have a transition from random ($R = 0$) to ordered ($R = 1$) layering with increasing depth, which is a feature of the Gulf Coast mudrocks. It is probable that the illitic I-S clays within the Brent

Group mudrocks were smectite-rich at the time of deposition, and have been progressively illitized during burial.

A single K-Ar age from a Brent Group mudrock of 108 Ma was significantly younger than the stratigraphic age, but older than K-Ar ages in interbedded sandstones (Matthews *et al.*, 1994). This was interpreted as a result of progressive illitization acting on a smectite-rich precursor mineral. Upper Jurassic mudrocks in the Central Graben include volcanic clays that are presently randomly ordered I-S with ~50% illite layers (Hansen & Lindgreen, 1989). This contrasts with the enclosing clastic sediment where the I-S has ~70–90% illite layers. The illitization of smectite within the volcanic sediments may be inhibited by lack of K (Ehrenberg, 1991; Chuhan *et al.*, 2000).

Mudrocks both within and above the Upper Jurassic Piper Formation (Inner Moray Firth) show progressive illitization of smectite, from ~40% smectite at 2400 m to practically pure illite at 4200 m (Burley & Flisch, 1989). K-Ar ages of <0.5 μm size fractions are between 153 and 68 Ma, with the oldest ages from the most shallow samples, interpreted as clay neoformation in Jurassic weathering profiles. Deeper illites have only slightly increased ^{40}Ar , but much increased K due to illitization, and consequently have much younger K-Ar ages. Burley & Flisch (1989) interpreted the data to indicate a two-stage illitization reaction within mudrocks: dissolution of smectite (with loss of ^{40}Ar) followed by neoformation of illite. Illitization was believed to have occurred 50–60 Ma ago.

Volumetrically, the dominant component of the clay fraction of the Upper Jurassic mudrocks is mixed-layer illite-smectite-vermiculite (I-S-V; Drits *et al.*, 1997), which has more usually been described as illite-smectite (I-S). Drits *et al.* (1997) recognized three groups of I-S-V. Group one I-S-V is predominantly detrital and has 0.69–0.73 illite, 0.26–0.20 smectite and 0.04–0.07 vermiculite interlayers, the illite, smectite and vermiculite interlayers being segregated. Group two I-S-V has been diagenetically transformed and has 0.80 illite, 0.12 smectite and 0.08 vermiculite interlayers, the vermiculite interlayers being segregated whereas the illite and smectite have the maximum ordering possible for $R = 1$. The I-S-V of group three has been further transformed during diagenesis and has 0.84 illite, 0.08 smectite and 0.08 vermiculite interlayers. The interlayer sites

within the I-S-V are partly filled with ammonium ions, the proportion of which increases from 0.03 to 0.20 per $O_{10}(OH)_2$ during burial from 1500–4000 m (Lindgreen *et al.*, 1991; Lindgreen, 1994). The end-member ammonium mineral is termed tobelite, and the ammonium is believed to be sourced from maturing organic material (Drits *et al.*, 2002). The ordering of the I-S changes during burial, from short-range ordered (IS) to long-range ordered (ISII; Lindgreen, 1994). The illitization reaction coincides with oil generation. Dypvik (1983) recorded discrete smectite as a component of shales within the Viking and Central Grabens. The smectite disappears at 65 to 75°C, presumably due to dissolution or illitization.

Lateral changes in shale mineralogy have been attributed to differences in detrital input, rather than to depth-related clay reactions (Shaw & Primmer, 1991). Lower Cretaceous–Upper Jurassic shale from Central Brae has smectite-rich, randomly interstratified I-S, whereas slightly deeper North Brae samples have illite-rich ordered I-S. However, the Central Brae mudrocks are very calcareous, suggesting that porefluids may have had high Ca^{2+}/K^+ , which would retard illitization (Eberl & Hower, 1976). Hence the observed changes in mineralogy could still be part of a general illitization of smectite. The ordered I-S was recorded as ISII or IS/ISII, and not the IS polytype. Shaw & Primmer (1991) record both authigenic illite and I-S (as well as kaolin). Recognisable authigenic clays are usually restricted to microfossil tests or secondary pores. Clays within open pores usually lack diagnostic fabrics, probably due to compaction.

Lindgreen *et al.* (1991) used a number of techniques to investigate the smectite-to-illite transition. As they found that illite layers have some smectite characteristics, they concluded that the transformation proceeded by a solid state mechanism. Studies by TEM of fundamental particle shape also throw light on growth mechanisms (Lindgreen *et al.*, 1992). The finest I-S separates include equal numbers of laths and equant particles (plates) at ~3500 m, but mostly equant particles at ~4500 m, suggesting that lath-like particles enlarge to become equant. Thickness distributions are dominated by 10 Å particles at both depths, though the deeper sample had significant numbers of 20–60 Å thick particles, suggesting particle thickening by the addition of 2:1 layers. These layers may develop from the outer edges of the particles, initially as isolated 'peaks', then as

rims. Rare spiral crystals observed at ~4500 m suggest that a spiral growth mechanism may also operate at relatively high temperatures.

Three K-Ar ages of shales from the Sleipner (or Pentland) Formation of the Brae area (western edge of the Viking Graben) were reported by Greenwood *et al.* (1994). Separated illite samples (115–84 Ma) were assumed to contain some detrital material, though this was not quantified. Following Shaw & Primmer (1991), Greenwood *et al.* (1994) considered that the illite in the mudrocks could be formed by direct precipitation, rather than being the end-point of the smectite-illite transformation. Stable oxygen isotope data of separated illites from the Heather Formation all lie close to equilibrium with Jurassic marine water at 60–80°C (Greenwood *et al.*, 1994). It seems unlikely that the connate marine porewaters within these mudrocks were ever displaced by externally sourced fluids.

Within the Kimmeridge Clay Formation are two turbiditic sandstone units, the Claymore Sandstone Member and the "Ten-foot" sandstone. In both units, authigenic clay minerals are rare, with only sparse rim cements of illite-smectite (Spark & Trewin, 1986).

CONCLUSIONS

(1) The clay minerals kaolin, illite and illite-smectite are common in North Sea oilfield sandstones: chlorite is only locally abundant. Smectite may have been common at the time of deposition, but has decreased in abundance with increased burial due to illitization.

(2) Kaolin originated by the dissolution of feldspars in meteoric porewater. Most UKCS sandstones have suffered this, including sandstones deposited on submarine fans. The only sands that may have escaped this process are the most distal offshore sands of the Fulmar Formation of the Central North Sea. Vermicular kaolin formed during early burial (surface ~40°C) recrystallized to blocky kaolin at 40–80°C in porewaters with a predominantly meteoric stable isotope signature. A relationship between $\delta^{18}O$ and δD exists for some kaolin suites, with depleted δD values suggesting direct interaction between clay minerals and organic matter.

(3) Early pore-lining illite has been interpreted as both infiltrated clastic clay, and as an early diagenetic phase. Early clays may have been smectite-rich illites, or even discrete smectites.

Later, fibrous illite is undoubtedly neoformed, and can degrade reservoir quality significantly. Within both sandstones and shales, there is an apparent increase in the K content over 4 km of burial, which could be due to swamping of the early smectite-rich phase by new growth illite, or to the progressive illitization of existing I-S. Much of the 'illite' that has been dated by the K-Ar method may actually be I-S. There is an accompanying change in crystal shape during burial, from laths to plates.

(4) The controls on the formation of authigenic illite are only poorly known, though temperature must play a role. Illite growth has been proposed for almost the entire range of diagenetic temperatures (e.g. 15–20°C Kantorowicz (1984); 35–40°C, Jeans & Atherton (1989) Oxfordian Sand, Inner Moray Firth; 50–90°C McLaughlin (1992) Brae Sst; 100–110°C Le Gallo *et al.* (1998); 130–140°C in Haltenbanken, Ehrenberg & Nadeau (1989)). It seems unlikely that there is a threshold temperature below which illite growth is impossible (or too slow to be significant), though this is a recurring hypothesis in the literature. Instead, illite growth seems to be an event, commonly triggered by oil emplacement or another change in the physico-chemical conditions within the sandstone such as an episode of overpressure release.

(5) While much is known about illite in the North Sea, there is much that remains unknown. Most of these unknowns are equally applicable to the illite from any sedimentary basin world wide, and must be viewed within this context. For example, K-Ar age dates are potentially vital for calibrating basin models, yet there is no consensus concerning the degree to which contamination affects the data. Promising avenues of research include detailed illite chemistry, or mineralogy.

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REFERENCES

- Barclay S.A. & Worden R.H. (2000) Geochemical modelling of diagenetic reactions in a sub-arkosic sandstone. *Clay Minerals*, **35**, 57–67.
- Beaufort D., Cassagnabere A., Petit S., Lanson B., Berger G., Lacharpagne J.C. & Johansen H. (1998) Kaolinite-to-dickite reaction in sandstone reservoirs. *Clay Minerals*, **33**, 297–316.
- Bjørkum P.A. & Gjelsvik N. (1988) An isochemical model for formation of authigenic kaolinite, K-feldspar and illite in sediments. *Journal of Sedimentary Petrology*, **58**, 506–511.
- Bjørkum P.A., Mjos R., Walderhaug O. & Hurst, A. (1990) The role of the late Cimmerian unconformity for the distribution of kaolinite in the Gullfaks field, northern North Sea. *Sedimentology*, **37**, 395–406.
- Bjørlykke K. & Brendsall A. (1986) Diagenesis of the Brent Group in the Staffjord Field. Pp. 157–168 in: *Roles of Organic Matter in Sedimentary Diagenesis* (D.L. Gaultier, editor). SEPM Special Publication, **38**. SEPM, Tulsa, Oklahoma.
- Bjørlykke K., Nedkvitne T., Ramm M. & Saigal G.C. (1992) Diagenetic processes in the Brent Group (Middle Jurassic) reservoirs, of the North Sea: an overview. Pp. 263–287 in: *Geology of the Brent Group* (A.C. Morton, R.S. Haszeldine, M.R. Giles & S. Brown, editors). Special Publication **61**. Geological Society, London.
- Blackbourn G.A. (1984) Diagenetic history and reservoir quality of a Brent sand sequence. *Clay Minerals*, **19**, 377–390.
- Blanche J.B. & Whitaker J.H.McD. (1978) Diagenesis of part of the Brent Sand Formation (Middle Jurassic) of the northern North Sea Basin. *Journal of the Geological Society of London*, **135**, 73–82.
- Brint J.F. (1989) *Isotope diagenesis and paleofluid movement: Middle Jurassic Brent sandstones, North Sea*. Unpublished PhD thesis, University of Strathclyde, UK.
- Burley S.D. (1986) The development and destruction of porosity within Upper Jurassic reservoir sandstones of the Piper and Tartan fields, Outer Moray Firth, North Sea. *Clay Minerals*, **21**, 649–694.
- Burley S.D. & Flisch M. (1989) K-Ar geochronology and the timing of detrital I/S clay illitization and authigenic illite precipitation in the Piper and Tartan Fields, Outer Moray Firth, UK North Sea. *Clay Minerals*, **24**, 285–315.
- Cavanagh A.J. (2002) *Oil and water-fibrous illite diagenesis and the onset of hydrocarbon charge in the petroleum systems of the Northern North Sea*. Unpublished PhD thesis, University of Edinburgh, UK.
- Chuhan F.A., Bjørlykke K. & Lowrey C. (2000) The role of provenance in illitization of deeply buried reservoir sandstones from Haltenbanken and north Viking Graben, offshore Norway. *Marine Petroleum Geology*, **17**, 673–689.
- Chuhan F.A., Bjørlykke K. & Lowrey C.J. (2001) Closed-system burial diagenesis in reservoir sandstones: Examples from the Garm Formation at Haltenbanken area, offshore mid-Norway. *Journal of Sedimentary Research*, **71**, 15–26.
- Clauer N. & Chaudhuri S. (1995) *Clays in Crustal*

- Environments, Isotope Dating and Tracing*. Springer Verlag Telos, Germany.
- Darby D., Wilkinson M., Haszeldine R.S. & Fallick A.E. (1997) Illite dates record deep fluid movements in petroleum basins. *Petroleum Geoscience*, **3**, 133–140.
- De'Ath N.G. & Schuyleman S.F. (1981) The geology of the Magnus oilfield. Pp. 342–351 in: *Petroleum Geology of the Continental Shelf of North-West Europe* (L.V. Illing & G.D. Hobson, editors). Institute of Petroleum, Heyden, London.
- Deer W.A., Howie R.A. & Zussman J. (1962) *Rock-forming Minerals, Volume 3, Sheet Silicates*. Longman, Essex, UK.
- Drits V.A., Sakharov B.A., Lindgreen H. & Salyn A. (1997) Sequential structure transformation of illite-smectite-vermiculite during diagenesis of Upper Jurassic shales from the North Sea and Denmark. *Clay Minerals*, **32**, 351–371.
- Drits V.A., Lindgreen H., Sakharov B.A., Jakobsen H.J., Salyn A.L. & Dainyak, L.G. (2002) Tobelization of smectite during oil generation in oil-source shales. Application to North Sea illite-tobelite-smectite-vermiculite. *Clays and Clay Minerals*, **50**, 82–98.
- Durand C., Matthews J.C., le Gallo Y., Brosse E. & Sommer F. (2000) Illitization and silicification in Greater Alwyn: I. Assessing and synthesizing experimental data. *Clay Minerals*, **35**, 211–225.
- Dypvik H. (1983) Clay mineral transformations in Tertiary and Mesozoic sediments from North Sea. *American Association of Petroleum Geologists Bulletin*, **67**, 160–165.
- Eberl D. & Hower J. (1976) Kinetics of illite formation. *Geological Society of America Bulletin*, **87**, 1326–1330.
- Egeberg P.K. & Aagaard P. (1989) Origins and evolution of formation waters from oil fields on the Norwegian Shelf. *Applied Geochemistry*, **4**, 131–142.
- Ehrenberg S.N. (1991) Kaolinized, K-leached zones at the contacts of the Garn Formation, Haltenbanken, mid-Norwegian continental shelf. *Marine Petroleum Geology*, **8**, 250–269.
- Ehrenberg S.N. (1993) Preservation of anomalously high-porosity in deeply buried sandstones by grain-coating chlorite – examples from the Norwegian Continental Shelf. *American Association of Petroleum Geologists Bulletin*, **77**, 1260–1286.
- Ehrenberg S.N. & Nadeau P.H. (1989) Formation of diagenetic illite in sandstones of the Garn Formation, Haltenbanken area, mid-Norwegian continental shelf. *Clay Minerals*, **24**, 233–253.
- Ehrenberg S.N., Aagaard P., Wilson M.J., Fraser A.R. & Duthie D.M.L. (1993) Depth-dependent transformation of kaolinite to dickite in sandstones of the Norwegian Continental Shelf. *Clay Minerals*, **28**, 325–352.
- Ehrenberg S.N., Dalland A., Nadeau P.H., Mearns E.W. & Amundsen H.E.F. (1998) Origin of chlorite enrichment and neodymium isotopic anomalies in Haltenbanken sandstones. *Marine Petroleum Geology*, **15**, 403–425.
- Emery D. & Robinson A. (1993) *Inorganic Geochemistry: Applications to Petroleum Geology*. Blackwell Scientific Publications, London.
- Emery D., Smalley P.C. & Oxtoby N.H. (1993) Synchronous oil migration and cementation demonstrated by a description of diagenesis. *Philosophical Transactions of the Royal Society A*, **334**, 115–125.
- Fairbridge R.W. (1967) Phases of diagenesis and authigenesis. Pp. 551 in: *Diagenesis in Sediments* (G. Larsen & G.V. Chilingar, editors). Developments in Sedimentology, **8**. Elsevier, Amsterdam.
- Fallick A.E., Macaulay C.I. & Haszeldine R.S. (1993) Implications of linearly correlated oxygen and hydrogen isotopic compositions for kaolinite and illite in the Magnus Sandstone, North Sea. *Clays and Clay Minerals*, **41**, 184–190.
- Faure G. (1986) *Principals of Isotope Geology*, 2nd edition. John Wiley and Sons, New York.
- Giles M.R. & de Boer R.B. (1990) Origin and significance of redistributional secondary porosity. *Marine Petroleum Geology*, **7**, 378–397.
- Giles M.R., Stevenson S., Martin S.V., Cannon S.J.C., Hamilton P.J., Marshall J.D. & Samways G.M. (1992) The reservoir properties and diagenesis of the Brent Group: a regional perspective. Pp. 289–327 in: *Geology of the Brent Group* (A.C. Morton, R.S. Haszeldine, M.R. Giles & S. Brown, editors). Special Publication, **61**. Geological Society London.
- Girard J.P., Munz I.A., Johansen H., Lacharpagne J.C. & Sommer F. (2002) Diagenesis of the Hild Brent sandstones, northern North Sea: isotopic evidence for the prevailing influence of deep basinal water. *Journal of Sedimentary Research*, **72**, 746–759.
- Glasmann J.R. (1992) The fate of feldspars in Brent Group Reservoirs, North Sea: a regional synthesis of diagenesis in shallow, intermediate and deep burial environments. Pp. 329–350 in: *Geology of the Brent Group* (A.C. Morton, R.S. Haszeldine, M.R. Giles & S. Brown, editors). Special Publication, **61**. Geological Society of London.
- Glasmann J.R., Clark R.A., Larter S., Briedis N.A. & Lundegard P.D. (1989a) Diagenesis and hydrocarbon accumulation, Brent Sandstone (Jurassic), Bergen High area, North Sea. *American Association of Petroleum Geologists Bulletin*, **73**, 1341–1360.
- Glasmann J.R., Larter S., Briedis N.A. & Lundegard P.D. (1989b) Shale diagenesis in the Bergen High area, North Sea. *Clays and Clay Minerals*, **37**, 97–112.
- Glasmann J.R., Lundegard P.D., Clark R.A., Penny B.K. & Collins I.D. (1989c) Geochemical evidence for the history of diagenesis and fluid migration: Brent

- Sandstone, Heather Field, North Sea. *Clay Minerals*, **24**, 255–284.
- Greenwood P.J., Shaw H.F. & Fallick A.E. (1994) Petrographic and isotopic evidence for diagenetic processes in Middle Jurassic sandstones and mudrocks from the Brae area, North Sea. *Clay Minerals*, **29**, 637–650.
- Hamilton P.J., Kelley S. & Fallick A.E. (1989) K-Ar dating of illite in hydrocarbon reservoirs. *Clay Minerals*, **24**, 215–232.
- Hamilton P.J., Giles M.R. & Ainsworth, P. (1992) K-Ar dating of illites in Brent Group reservoirs: a regional perspective. Pp. 377–400 in: *Geology of the Brent Group* (A.C. Morton, R.S. Haszeldine, M.R. Giles & S. Brown, editors). Special Publication, **61**. Geological Society of London.
- Hancock N.J. & Taylor A.M. (1978) Clay mineral diagenesis and oil migration in the Middle Jurassic Brent Sand Formation. *Journal of the Geological Society of London*, **135**, 69–72.
- Hansen P.L. & Lindgreen H. (1989) Mixed-layer illite/smectite diagenesis in Upper Jurassic claystones from the North Sea and onshore Denmark. *Clay Minerals*, **24**, 197–213.
- Harris N.B. (1992) Burial diagenesis of Brent sandstones: a study of Staffjord, Hutton and Lyell fields. Pp. 351–375 in: *Geology of the Brent Group* (A.C. Morton, R.S. Haszeldine, M.R. Giles & S. Brown, editors). Special Publication, **61**. Geological Society of London.
- Hassouta L. (1999) *La comparaison de grès cimentés et de grès non cimentés par la calcite du groupe du Brent (Zone d'alwyn, Mer du Nord). Une clé pour l'établissement de bilans de matière et la compréhension des processus de formation du quartz et des argilles (illite, kaolinite, dickite)*. PhD thesis, L'Université des Sciences et Technologies de Lille, France.
- Hassouta L., Buatier M.D., Potdevin J.L. & Liewig N. (1999) Clay diagenesis in the sandstone reservoir of the Ellon Field (Alwyn, North Sea). *Clay Minerals*, **47**, 269–285.
- Haszeldine R.S., Brint J.F., Fallick A.E., Hamilton P.J. & Brown S. (1992) Open and restricted hydrologies in Brent Group diagenesis: North Sea. Pp. 401–419 in: *Geology of the Brent Group* (A.C. Morton, R.S. Haszeldine, M.R. Giles & S. Brown, editors). Special Publication, **61**. Geological Society of London.
- Haszeldine R.S., Wilkinson M., Derby D., Macaulay C.I., Couples G.C., Fallick A.E., Fleming C.G., Stewart R.N.T. & McAulay G. (1999) Diagenetic porosity creation in an overpressured graben. Pp. 1339–1350 in: *Petroleum Geology of NW Europe* (S. Boldy & S. Fleet, editors). Geological Society of London.
- Heavyside J., Langley G.O., & Pallatt N. (1983) Permeability characteristics of Magnus reservoir rock. *8th SPWLA London Chapter European Formation Evaluation Symposium Transactions*, London, March, 1983, Paper A.
- Hendry J.P. & Trewin N.H. (1995) Authigenic quartz microfabrics in Cretaceous turbidites – evidence for silica transformation processes in sandstones. *Journal of Sedimentary Research*, **A65**, 380–392.
- Hendry J.P., Wilkinson M., Fallick A.E. & Trewin N.H. (2000) Disseminated 'jigsaw-piece' dolomite in Upper Jurassic shelf sandstones, Central North Sea: an example of cement growth during bioturbation? *Sedimentology*, **47**, 631–644.
- Hillier S. (1994) Pore-lining chlorites in siliciclastic reservoir sandstones – electron-microprobe, SEM and XRD data, and implications for their origin. *Clay Minerals*, **29**, 665–679.
- Hogg A.J.C. (1989) *Petrographic and isotopic constraints on the diagenesis and reservoir properties of the Brent Group sandstones*. PhD thesis, University of Aberdeen, UK.
- Hogg A.J.C., Pearson M.J., Fallick A.E., Hamilton P.J. & Macintyre R. (1987) Clay mineral and isotope evidence for control on reservoir properties of Brent Group sandstones, British North Sea. *Terra Incognita*, **7**, 342.
- Hogg A.J.C., Hamilton P.J. & Macintyre R.M. (1993) Mapping diagenetic fluid flow within a reservoir: K-Ar dating in the Alwyn area (UK North Sea). *Marine Petroleum Geology*, **10**, 279–294.
- Hogg A.J.C., Pearson M.J., Fallick A.E. & Hamilton P.J. (1995) An integrated thermal and isotopic study of the diagenesis of the Brent Group, Alwyn South, UK North Sea. *Applied Geochemistry*, **10**, 531–546.
- Hower J., Eslinger E.V., Hower M.E. & Perry, E.A. (1976) Mechanism of burial metamorphism of argillaceous sediment: 1. Mineralogical and chemical evidence. *Geological Society of America Bulletin*, **87**, 725–737.
- Huang W.L., Bishop A.M. & Brown R.W. (1986) The effect of fluid/rock ratio on feldspar dissolution and illite formation under reservoir conditions. *Clay Minerals*, **21**, 585–601.
- Hurst A. & Irwin H. (1982) Geological modelling of clay diagenesis in sandstones. *Clay Minerals*, **17**, 5–22.
- Jahren J.S. & Aagaard P. (1989) Compositional variations in diagenetic chlorites and illites, and relationships with porewater chemistry. *Clay Minerals*, **24**, 157–170.
- Jeans C.V. & Atherton A.F. (1989) Silicate and associated cements in an Oxfordian marine-freshwater transition, Inner Moray Firth, UK North Sea. *Clay Minerals*, **24**, 317–339.
- Jeans C.V. & Fisher M.J. (1986) Diagenesis in Upper Jurassic marine sandstones from the North Sea well 14/26-1 and its significance. *Clay Minerals*, **21**, 513–535.

- Jeremiah J.M. & Nicholson P.H. (1999) Middle Oxfordian to Volgian sequence stratigraphy of the Greater Shearwater area. Pp. 153–170 in: *Petroleum Geology of Northwestern Europe: Proceedings of the 5th Conference* (A.J. Fleet & S.A.R. Boldy, editors). Geological Society, London.
- Jones T.G.J., Hughes T.L. & Tomkins P. (1989) The ion content and mineralogy of a North Sea Cretaceous shale formation. *Clay Minerals*, **24**, 393–410.
- Jordan A., Thomas M., Brevart O., Robson P., Sommer F. & Sullivan M. (1987) Diagenesis as the control of the Brent sandstone reservoir in the Greater Alwyn area (East Shetland Basin). Pp. 951–961 in: *Petroleum Geology of N.W. Europe* (J. Brooks & K.W. Glennie, editors). Graham & Trotman, London.
- Kantorowicz J. (1984) The nature, origin and distribution of authigenic Clay Minerals, from Middle Jurassic Ravenscar and Brent Group sandstones. *Clay Minerals*, **19**, 359–375.
- Kantorowicz J.D. (1990) The influence of variations in illite morphology on the permeability of Middle Jurassic Brent Group sandstones, Cormorant Field, UK North-Sea. *Marine Petroleum Geology*, **7**, 66–74.
- Land L.S. (1983) The application of stable isotopes to studies of the origin of dolomite and to problems of diagenesis of clastic sediments. Pp. 4.1–4.22 in: *Stable Isotopes in Sedimentary Geology* (M.A. Arthur, editor). SEPM Short Course **10**, SEPM, Tulsa, Oklahoma.
- Land L.S. & Dutton S.P. (1978) Cementation of a Pennsylvanian deltaic sandstone: isotopic data. *Journal of Sedimentary Petrology*, **48**, 1167–1176.
- Le Gallo Y., Bildstein O. & Brosse E. (1998) Coupled reaction-flow modeling of diagenetic changes in reservoir permeability, porosity and mineral compositions. *Journal of Hydrology*, **209**, 366–388.
- Liewig N., Clauer N. & Sommer F. (1987) Rb-Sr and K-Ar dating of clay diagenesis in a Jurassic sandstone reservoir. *Bulletin of the American Association of Petroleum Geologists*, **71**, 1467–1474.
- Lindgreen H. (1994) Ammonium fixation during illite-smectite diagenesis in upper Jurassic shale, North Sea. *Clay Minerals*, **29**, 527–537.
- Lindgreen H., Drits V.A., Sakharov B.A., Jakobsen H.J., Salyn A.L., Dainyak L.G. & Kroyer H. (2002) The structure and diagenetic transformation of illite-smectite and chlorite-smectite from North Sea Cretaceous-Tertiary chalk. *Clay Minerals*, **37**, 429–450.
- Lindgreen H., Jacobsen H. & Jakobsen H.J. (1991) Diagenetic structural transformations in North Sea Jurassic illite/smectite. *Clays and Clay Minerals*, **39**, 54–69.
- Lindgreen H., Garnaes J., Besenbacher F., Laegsgaard E. & Stensgaard I. (1992) Illite-smectite from the North-Sea investigated by scanning tunnelling microscopy. *Clay Minerals*, **27**, 331–342.
- Lønøy A., Akselsen J. & Rønning K. (1986) Diagenesis of a deeply buried sandstone reservoir: Hild Field, Northern North Sea. *Clay Minerals*, **21**, 497–512.
- Macaulay C.I. (1990) *Clastic diagenesis and porefluid evolution: an isotopic study, Magnus Oilfield, North Sea*. Unpublished PhD thesis, University of Strathclyde, UK.
- Macaulay C.I., Haszeldine R.S. & Fallick A.E. (1992) Diagenetic pore waters stratified for at least 35 million years: Magnus oil field, North Sea. *Bulletin of the American Association of Petroleum Geologists*, **76**, 1625–1634.
- Maliva R.G., Dickson J.A.D. & Fallick A.E. (1999) Kaolin cements in limestones: potential indicators of organic-rich porewaters during diagenesis. *Journal of Sedimentary Research*, **69**, 158–163.
- Matthews J.C., Velde B. & Johansen H. (1994) Significance of K-Ar ages of authigenic illitic clay-minerals in sandstones and shales from the North Sea. *Clay Minerals*, **29**, 379–389.
- McAulay G.E., Burley S.D. & Johns L.H. (1993) Silicate mineral authigenesis in the Hutton and NW Hutton fields: implications for sub-surface porosity development. Pp. 1377–1394 in: *Petroleum Geology of Northwest Europe: Proceedings of the 4th Conference* (J.R. Parker, editor). Geological Society, London.
- McAulay G.E., Burley S.D., Fallick A.E. & Kusznir N.J. (1994) Paleohydrodynamic fluid flow regimes during diagenesis of the Brent Group in the Hutton-Northwest Hutton reservoirs: constraints from oxygen isotope studies of authigenic kaolin and reverse flexural modelling. *Clay Minerals*, **29**, 609–626.
- McBride J.J. (1992) *The diagenesis of middle Jurassic reservoir sandstones of Bruce Field, UK, North Sea*. Unpublished PhD thesis, University of Aberdeen, UK.
- McHardy W.J., Wilson M.J. & Tait J.M. (1982) Electron microscope and X-ray diffraction studies of filamentous illitic clay from sandstones of the Magnus Field. *Clay Minerals*, **17**, 23–29.
- McLaughlin Ó.M. (1992) *Isotopic and textural evidence for diagenetic fluid mixing in the South Brae oilfield, North Sea*. Unpublished PhD thesis, University of Glasgow, UK.
- McLaughlin Ó.M., Haszeldine R.S., Fallick A.E. & Rogers, G. (1994) The case of the missing clay, aluminium loss and secondary porosity, South Brae Oilfield, North Sea. *Clay Minerals*, **29**, 651–664.
- Midtbo R.E.A., Rykkje J.M. & Ramm M. (2000) Deep burial diagenesis and reservoir quality along the eastern flank of the Viking Graben. Evidence for illitisation and quartz cementation after hydrocarbon emplacement. *Clay Minerals*, **35**, 227–237.
- Nadeau P.H. (1985) The physical dimensions of

- fundamental clay particles. *Clay Minerals*, **20**, 499–514.
- Nadeau P.H., Wilson M.J., McHardy W.J. & Tait, J.M. (1994) Interstratified clays as fundamental particles. *Science*, **225**, 923–925.
- Newman A.C.D. & Brown G. (1987) The chemical constitution of clays. Pp. 1–128 in: *Chemistry of Clays and Clay Minerals* (A.C.D. Newman, editor). Mineralogical Society, London.
- Osborne M., Haszeldine R.S. & Fallick A.E. (1994) Variation in kaolin morphology with growth temperature in isotopically mixed pore-fluids, Brent Group, UK North Sea. *Clay Minerals*, **29**, 591–608.
- Pearson M.J. & Small J.S. (1988) Illite-smectite diagenesis and palaeotemperatures in northern North Sea Quaternary to Mesozoic shale sequences. *Clay Minerals*, **23**, 109–132.
- Pearson M.J., Watkins D. & Small J.S. (1982) Clay diagenesis and organic maturation in Northern North Sea Sediments. Pp. 665–675 in: *Proceedings of the International Clay Conference, Pavia & Bologna* (H. Van Olphen & F. Veniale, editors). Developments in Sedimentology, **35**. Elsevier, Amsterdam.
- Pearson M.J., Watkins D., Pittion J.-L., Caston D. & Small J.S. (1983) Aspects of burial diagenesis, organic maturation and palaeothermal history of an area in the South Viking Graben, North Sea. Pp. 161–173 in: *Petroleum Geochemistry and Exploration of Europe* (J. Brooks, editor). Special Publication, **12**, Geological Society, London.
- Purvis K. (1995) Diagenesis of Lower Jurassic sandstones, block-211/13 (Penguin area), UK northern North Sea. *Marine Petroleum Geology*, **12**, 219–228.
- Raine S.C.R. (1987) *Sedimentology, diagenesis and geochemistry of the Magnus Sandstone Member, Northern North Sea*. Unpublished PhD thesis, University of Edinburgh, UK.
- Ramm M. & Ryseth A.E. (1996) Reservoir quality and burial diagenesis in the Statfjord Formation, North Sea. *Petroleum Geoscience*, **2**, 313–324.
- Scotchman I.C., Johns L.H. & Miller R.S. (1989) Clay diagenesis and oil migration in Brent Group sandstones of NW Hutton Field, UK North Sea. *Clay Minerals*, **24**, 339–374.
- Shaw H.F. & Primmer T.J. (1991) Diagenesis of mudrocks from the Kimmeridge Clay Formation of the Brae Area, UK North Sea. *Marine Petroleum Geology*, **8**, 270–277.
- Small J.S., Hamilton D.L. & Habesch S. (1992) Experimental simulation of clay precipitation within reservoir sandstones. 2: Mechanism of illite formation and controls on morphology. *Journal of Sedimentary Petrology*, **62**, 520–529.
- Sommer F. (1978) Diagenesis of Jurassic sandstones in the Central Graben. *Journal of the Geological Society of London*, **135**, 63–67.
- Spark I.S.C. & Trewin N.H. (1986) Facies-related diagenesis in the Main Claymore Oilfield sandstones. *Clay Minerals*, **21**, 479–496.
- Stewart D.J. (1986) Diagenesis of the shallow marine Fulmar Formation in the Central North Sea. *Clay Minerals*, **21**, 537–564.
- Swarbrick R.E. (1994) Reservoir diagenesis hydrocarbon migration under hydrostatic palaeopressure conditions. *Clay Minerals*, **29**, 463–473.
- Taylor D.J. & Dietvorst J.P.A. (1991) The Cormorant Field, Blocks 211/21a, 211/26a, UK North Sea. Pp. 73–81 in: *United Kingdom Oil and Gas Fields 25 Years Commemorative Volume* (I.L. Abbotts, editor). Memoir, **14**. Geological Society, London.
- Thomas M. (1986) Diagenetic sequences and K-Ar dating in Jurassic sandstones, central Viking Graben: effects on reservoir properties. *Clay Minerals*, **21**, 695–710.
- Velde B. (1977) A proposed phase diagram for illite, expanding chlorite, corrensite and illite-montmorillonite mixed layer minerals. *Clays and Clay Minerals*, **25**, 264–270.
- Velde B. & Nicot E. (1985) Diagenetic clay mineral composition as a function of pressure, temperature, and chemical activity. *Journal of Sedimentary Petrology*, **55**, 541–547.
- Watson R.S. (1993) *The diagenesis of Tertiary sands from the Forth & Balmoral Fields, Northern North Sea*. Unpublished PhD thesis, University of Aberdeen, UK.
- Watson R.S., Trewin N.H. & Fallick A.E. (1995) The formation of carbonate cements in the Forth & Balmoral Fields, northern North Sea; a case for biodegradation, carbonate cementation and oil leakage during early burial. Pp. 177–200 in: *Characterisation of Deep Marine Clastic Systems* (A.J. Hartley & D.J. Prosser, editors). Special Publication, **94**. Geological Society, London.
- Wensaas L., Shaw H.F., Gibbon S.K., Aagaard P. & Dypvik H. (1994) Nature and causes of overpressuring in mudrocks of the Gullfaks area, North Sea. *Clay Minerals*, **29**, 439–449.
- Wilkinson M. & Haszeldine R.S. (2002a) Fibrous illite in oilfield sandstones – a nucleation kinetic theory of growth. *Terra Nova*, **14**, 56–60.
- Wilkinson M. & Haszeldine R.S. (2002b) Problems with argon: K-Ar ages in Gulf Coast shales. *Chemical Geology*, **191**, 277–283.
- Wilkinson M., Fallick A.E., Keaney G.M.J., Haszeldine R.S. & McHardy W. (1994) Stable isotopes in illite: the case for meteoric water flushing within the Upper Jurassic Fulmar Formation sandstones, UK North Sea. *Clay Minerals*, **29**, 567–574.
- Wilkinson M., Haszeldine R.S. & Fallick A.E. (2004) Hydrocarbon filling history from diagenetic evidence: Brent Group, UK North Sea. *Marine Petroleum Geology*, **21**, 443–455.

- Worden R.H. & Morad S. (2003) Clay minerals in sandstones: controls on formation, distribution and evolution. Pp. 3–42 in: *Clay Mineral Cements in Sandstones* (R.H. Worden & S. Morad, editors). Special Publication, **34**. International Association of Sedimentologists.