

Morphological development of basin-scale silica diagenetic fronts revealed with 2D seismic reflection data: offshore Sakhalin, Russian Far East

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Abstract: Two-dimensional (2D) seismic reflection data reveal two silica diagenetic boundaries located in the North Sakhalin Basin, offshore Sakhalin Island. These boundaries form distinctive high-amplitude seismic reflections that predominantly cross-cut Miocene stratigraphy. The boundaries represent the opal-A to opal-CT (cristobalite and tridymite) transition and the subsequent deeper opal-CT to quartz transition and cover an area of c. 107 000 km². The reflections represent diagenetic reaction fronts that advance through silica-rich host sediments during burial. The oldest strata within which the diagenetic fronts have been identified are of Miocene age and the fronts commonly show a parallel relationship to a Late Miocene unconformity. These diagenetic fronts develop a variety of morphological forms. In the North Sakhalin Basin these front geometries are probably the result of the interference of the diagenetic fronts with inclined and folded stratigraphy together with lateral changes in the composition and, therefore, depth of transformation of the sediment. We define criteria for the recognition of these 2D cross-sectional morphological elements and by so doing provide the basis for the objective description and categorization of diagenetic front geometry.

Biogenic silica (opal-A) is an important constituent of Neogene-age marine sediments, particularly in high-latitude and equatorial regions. The transformation of opal-A to opal-CT and subsequent transition of opal-CT to quartz are important diagenetic reactions that occur in siliceous sediments within marine sedimentary basins (Hesse 1990). Opal-A is extensively deposited in many types of depositional environments and diagenetic silica transformations have been identified from a number of sedimentary basins worldwide (e.g. Hein *et al.* 1978; Pisciotto 1981; Nobes *et al.* 1992a; Davies & Cartwright 2002; Volpi *et al.* 2003).

The opal-A to opal-CT conversion usually occurs within the first 1000 m of burial and changes the sediments' physical properties, such as acoustic velocity and density (e.g. Guerin & Goldberg 1996). This conversion is commonly recognized on seismic reflection data. Opal-CT is subsequently converted to quartz and this can also give rise to a seismic reflection, albeit generally less obvious on seismic data. Therefore, these silica transition boundaries (herein termed 'diagenetic fronts') can often be imaged with seismic reflection data and mapped over areas of up to 10⁵ km². Recent research has found that diagenetic fronts can develop complex and variable 3D forms (e.g. Davies 2005). Furthermore, the reaction is increasingly being linked to other important geological phenomena in sedimentary basins, such as (1) fluid flux (Eichhubl & Boles 1998), (2) the formation of clastic injectites (Davies *et al.* 2006), (3) submarine slope failures (Davies & Clark 2006) and (4) the formation of differential compaction folds above the boundaries (Davies 2005). These are all caused by diagenetic front advancement causing sediment volume change and fluid flow. However, little is known about how diagenetic fronts advance at kilometre to basin scale, and the role of and interplay between chemical and physical processes (e.g. hydrofracturing) is uncertain.

The aim of this paper is to describe an array of new and recently described opal-A to opal-CT diagenetic front geometries and to give the first descriptions of an opal-CT to quartz boundary using 2D seismic reflection data from offshore Sakhalin, far eastern Russia. Objective criteria for the cross-sectional

recognition of the variety of morphological forms are defined for the first time, to allow for a standard categorization of frontal geometries in different sedimentary basins worldwide. The paper then explores the potential linked chemical and physical processes that could account for the formation of the geometries, to further our understanding of the controls on large-scale diagenetic front advancement. The diagenetic boundaries identified within the North Sakhalin Basin have different morphologies from those previously described in other basins, such as the Faeroe–Shetland Basin (e.g. Davies 2005). The key features of the two contrasting sets of morphologies are compared and contrasted to gain a greater understanding of the mechanisms of conversion and advancement of the diagenetic boundaries.

Silica diagenetic fronts

The silica diagenetic reactions are dissolution–reprecipitation phase changes (Kastner *et al.* 1977; Williams & Crerar 1985; Williams *et al.* 1985; Tada 1991), which are primarily controlled by temperature and time (Keller & Isaacs 1985; Hesse 1990; Tada 1991; Nobes *et al.* 1992a). However, host rock lithology, together with the pore water chemistry, also play an important role (Kastner *et al.* 1977; Isaacs 1981; Hinman 1990; Tada 1991). A combination of the above factors could lead to some transformations taking place at shallower burial depths and lower temperatures.

The temperatures required for opal-A to opal-CT transformation can be highly variable, ranging from 2 to 56 °C (Hein *et al.* 1978; Keller & Isaacs 1985; Nobes *et al.* 1992a; Bohrmann *et al.* 1994). Temperatures needed for the opal-CT to quartz transition range from 46 to 110 °C (Hesse 1990; Nobes *et al.* 1992a). The depths at which the transitions occur vary and depend on the temperature at which conversion is initiated. Temperature is one of the dominant controls on these reactions, therefore the opal-A to opal-CT front has been considered to mark isotherms (Hein *et al.* 1978; Kuramoto *et al.* 1992; Nobes *et al.* 1992b; Lee *et al.* 2003) that can in some cases be parallel

to the sea bed. Both of the silica phase change reactions are reliant on a critical temperature range to initiate conversion; therefore the two boundaries are commonly parallel to one another. Relatively speaking, transition zones advance upwards through the sedimentary section, or one could think of it as the sediment subsiding through the transition, which is located at a specific depth for any one lithology type, termed the depth of conversion (DOC; Davies 2005). The transition zones are reaction fronts, which are defined as a layer of reaction that propagates through a chemical mixture, converting the initial reactants to final products (Scott 1994). Reaction fronts and the processes by which they form are widely studied, as they are found in chemical, biological and geological systems.

The majority of siliceous sediments have extremely high porosity as a result of abundant interskeletal pores. The conversion of opal-A to opal-CT causes a dramatic reduction of intra- and inter-skeletal porosity (*c.* 25–35% decrease) allowing sediments to consolidate at depth (Langseth & Tamaki 1992; Chaika & Dvorkin 2000; Volpi *et al.* 2003). The opal-CT to quartz transition results in a further, less dramatic, reduction, as the remaining porosity is lost by chemical compaction (e.g. pressure-solution). The porosity reduction and associated fluid expulsion that occurs during the conversion can lead to the generation of shallow overpressure; when the pressure is released the resulting fluid flow towards the sediment surface can cause a variety of deformational and compactional features (Volpi *et al.* 2003). The changes in the physical properties of the sediment give rise to normal (positive) polarity, moderate- to high-amplitude seismic reflections and common amplitude terminations that can cross-cut stratigraphy (e.g. Davies & Cartwright 2002).

Geological setting and database

Sakhalin Island is located along the western margin of the Sea of Okhotsk (Fig. 1). The Sea of Okhotsk is a relatively shallow marginal sea forming the northwestern arm of the Pacific Ocean. The sea has an average depth of *c.* 820 m, reaching *c.* 3400 m at its deepest point near the Kuril Islands. It is an area where biogenic siliceous sediments accumulated during the Miocene, which were overlain by deltaic sediments during the Pliocene (Tull 1997). The lithologies in the basin consist of interbedded diatomites and diatomaceous mudstones with some sandstones. Diatomites are rocks chiefly composed of the opal-A tests of diatoms, whereas diatomaceous mudstones are clay-rich sediments that still have significant opal-A content in the form of diatoms.

The study area is located within the North Sakhalin Basin, which is a deep (up to 12 km), Tertiary strike-slip downwarp associated with a major north–south-trending dextral strike-slip fault system, known as the Sakhalin–Hokkaido Dextral Shear Zone (Fig. 1), that formed as a result of Neogene displacement (Khvedchuk 1993; Worrall *et al.* 1996; Lindquist 2000; Weaver *et al.* 2004). The basin fill unconformably overlies a Cretaceous to Palaeocene basement accretionary complex. This in turn is overlain by Oligocene synrift deposits and Miocene diatomaceous siliciclastic marine sediments, together with thick successions of deltaic sediments that accumulated in the Late Miocene–Pliocene (Tull 1997). These deltaic sediments are thought to have been deposited by the eastward-prograding palaeo-Amur River system (Lindquist 2000). Unreleased well data indicate that the basin has a geothermal gradient of 35–40 °C km⁻¹.

The diagenetic fronts are hosted within Miocene and Pliocene

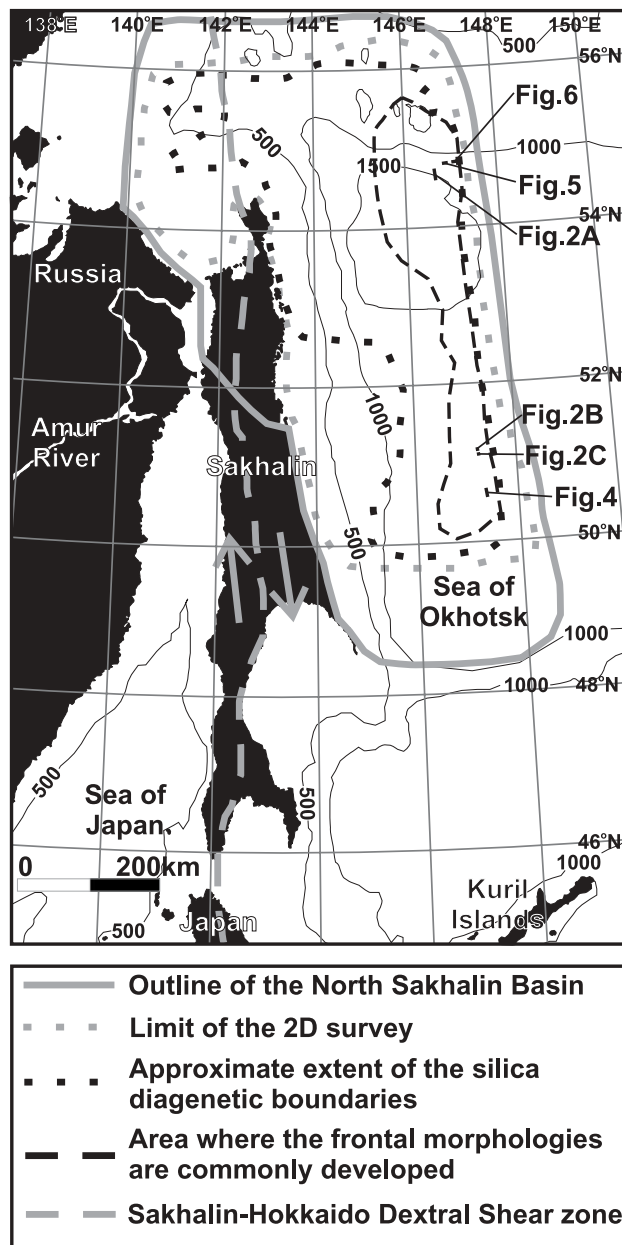


Fig. 1. Map of Sakhalin showing the location of the North Sakhalin Basin. Bathymetry (fine black lines) shown in metres. Locations of subsequent figures that use images from the seismic dataset are indicated.

sediments. These sediments are deformed as a result of a phase of Late Miocene folding. There was also an episode of Late Pliocene inversion that resulted in folding and thrusting, the modification of pre-existing structures and in significant amounts of erosion (Tull 1997). The folds are oriented NW–SE and have a wavelength of 0.5–2 km.

The study area is covered by both a 2D (Fig. 1) and a 3D seismic survey. The 2D survey, acquired in 2004, consists of 7600 km of 2D multi-channel seismic reflection data, with 13 km × 13 km line spacing and 25 m shot interval. The 2D survey has a seismic resolution of *c.* 10 m. Both surveys image the diagenetic fronts; however, it is the 2D survey that provides the best images for interpretation and is used in this paper. This

is probably because the 2D survey images a larger part of the basin; hence providing more examples of the front geometry. Seismic character alone is insufficient to constrain the origin of the high-amplitude, cross-cutting reflection events. However, lithological calibration of the seismic data is provided by off-shore drilling, which indicates that these reflections are coincident with the transitions of opal-A to opal-CT and of opal-CT to quartz. To convert from two-way travel time to depth we have assumed an average seismic velocity of 2100 m s^{-1} for sediment below the opal-A to opal-CT transition and 1800 m s^{-1} for sediment above this diagenetic front. These velocities are consistent for converted and unconverted sediment (i.e. opal-A-bearing sediment prior to conversion to opal-CT) at these burial depths from other documented examples worldwide (e.g. Guerin & Goldberg 1996).

General characteristics

The diagenetic fronts are not observed in the Cretaceous basement or in Oligocene-age sediments, and where the basement forms fault blocks the boundaries terminate. The top of the Miocene succession is marked by an unconformity (herein termed the Late Miocene Unconformity; Fig. 2a), represented by a high-amplitude, continuous zero-phase reflection with a positive acoustic impedance contrast. The diagenetic boundaries are generally observed to cross-cut the Miocene succession, but the opal-A to opal-CT front cross-cuts the Late Miocene Unconfor-

mity and passes into Pliocene sediments in a few areas of the basin.

We identify two seismic boundaries generally separated by 300–400 m that cover an area of $c. 107\,000 \text{ km}^2$. The interpretation of the boundaries as silica transitions is further supported by the positive polarity and high amplitude of the two reflections (Fig. 2a), which are consistent with them being diagenetic boundaries where density and acoustic velocity increase. The upper reflection is also very similar to an opal-A to opal-CT boundary in the Faeroe–Shetland Basin proven by drilling (e.g. Davies & Cartwright 2002). Therefore in the remainder of this paper the upper seismic reflection is termed the opal-A to opal-CT boundary and the lower reflection is termed the opal-CT to quartz boundary.

The opal-CT to quartz boundary is generally of lower amplitude than the opal-A to opal-CT boundary (Fig. 2a). The boundaries are generally parallel to each other, with only localized variations in this relationship. Regionally, the opal-A to opal-CT boundary occurs at $c. 500\text{--}1400 \text{ m}$ and the opal-CT to quartz boundary occurs at $c. 800\text{--}2400 \text{ m}$ below the present-day sea bed. These depths of transformation correspond to temperatures of $c. 17.5\text{--}56^\circ\text{C}$ for the opal-A to opal-CT transition and $c. 28\text{--}96^\circ\text{C}$ for the opal-CT to quartz transition (assuming a geothermal gradient of $35\text{--}40^\circ\text{C km}^{-1}$). The shallowest occurrences relative to the sea bed are found along the margins and in the nearshore parts of the North Sakhalin Basin. The fronts are mainly concordant with stratigraphy, but cross-cut deformed and inclined strata (Fig. 2a, 1).

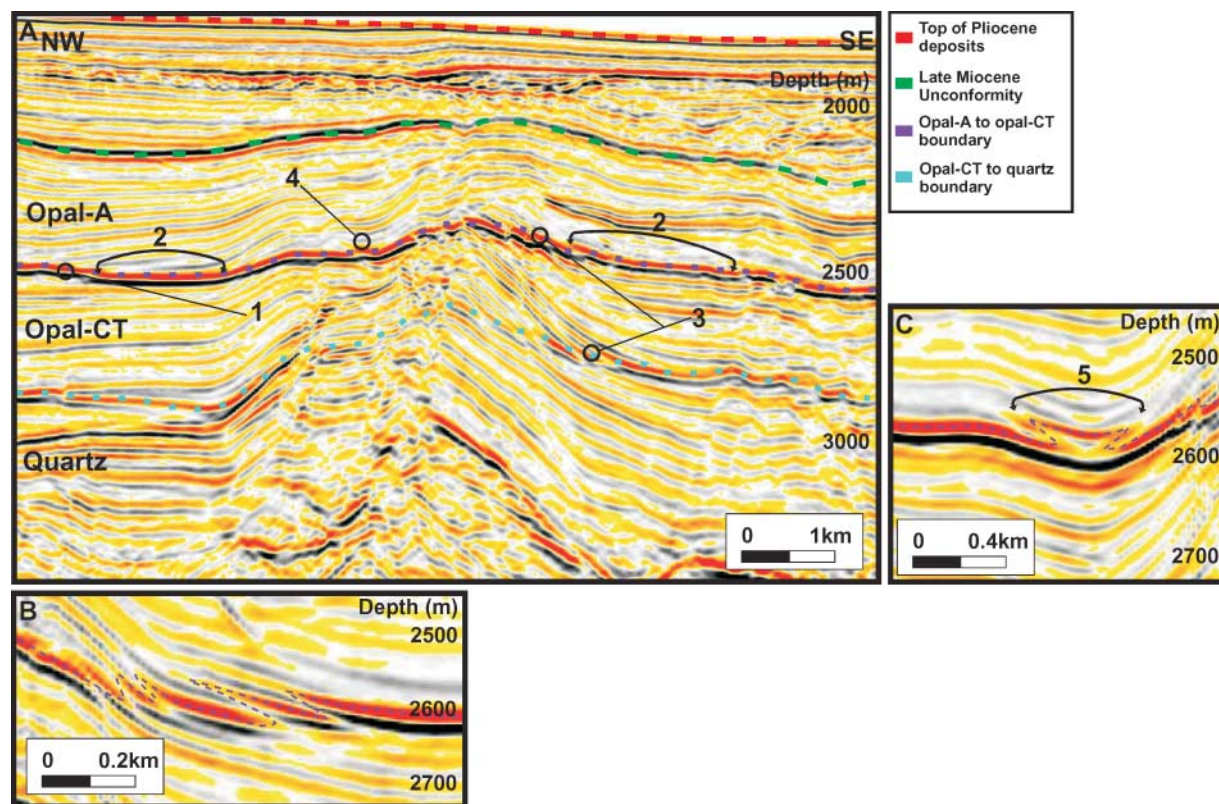


Fig. 2. (a) Seismic line showing the characteristics of the silica diagenetic fronts. 1, Cross-cutting of inclined strata; 2, planar fronts; 3, serrated pattern; 4, attached wing. The same horizon colour convention is used throughout this paper. (b) Close-up of a serrated pattern in the opal-A to opal-CT front. (Note the sawtooth shape of the front as it cross-cuts inclined stratigraphy.) (c) Close-up of an attached wing in the opal-A to opal-CT front. (Note the concave shape and the upward tapering edges of the feature, and the differential subsidence (5) of the wing's centre and of the overlying sediment.)

The diagenetic fronts are not parallel to the sea bed, unlike many other silica transitions (e.g. Hein *et al.* 1978). Instead, in most areas they are parallel to the Late Miocene Unconformity (Fig. 2a). Furthermore, the diagenetic fronts are not obvious on all 2D seismic lines and on some they occur only across part of the line.

Diagenetic front morphologies

We adopt some of the nomenclature proposed by Davies & Cartwright (2002) and Davies (2005) and expand upon it, describing and providing objective criteria for the recognition of several new morphological features. These morphological features are almost always better developed on the opal-A to opal-CT front than on the opal-CT to quartz front. In the following description we start by describing the simplest front morphologies before analysing the more complex forms. The cross-sectional geometry of host strata above and below the diagenetic front and the cross-sectional form of the fronts allow for a simple and objective nomenclature to be proposed for the North Sakhalin Basin examples. For this classification to be useful and

objective we have devised simple recognition criteria. Eight categories are defined here (Fig. 3) but we expect this could be further developed by seismic imaging of other sedimentary basins where opal-A-rich sediments accumulate.

Distribution of diagenetic front morphologies

The frontal morphologies associated with the silica diagenetic boundaries occur throughout the basin fill where these boundaries have been identified. However, the morphologies are more commonly developed along the easternmost margin of the North Sakhalin Basin (Fig. 1); that is, the more distal and deeper marine areas. Multi-kilometre-scale depressions and elevations (see below) are almost exclusively formed in this area of the basin.

Planar fronts (Fig. 3a)

Planar fronts are featureless regions of the diagenetic fronts that are located between other morphological features (Fig. 2a, 2). They can be planar over multi-kilometre distances. They can be

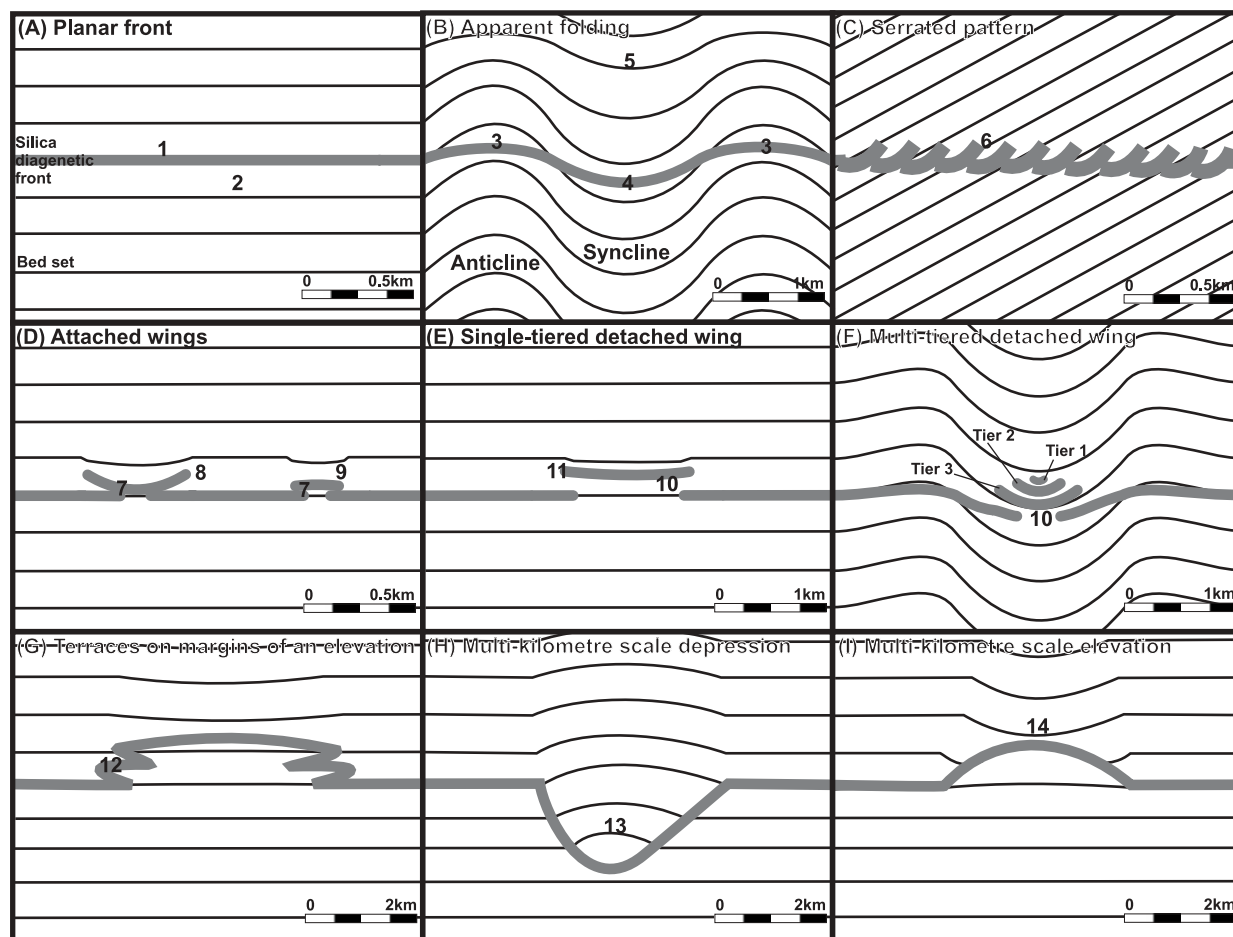


Fig. 3. Schematic diagram summarizing the key seismic attributes of the various morphological features (a–i). 1, Lack of frontal relief; 2, track stratigraphy; 3, pseudo-antiform; 4, pseudo-synform; 5, front is parallel to a higher stratal reflection; 6, up-dip sawtooth pattern of front as it cross-cuts inclined stratigraphy; 7, base attached to main diagenetic front; 8, concave-upward shape with tapering margins; 9, flat to dome shape; 10, base detached from main diagenetic front; 11, slight concave-upward to elongate shape; 12, terrace; 13, gentle antiformal folding; 14, downward flexure of overlying sediment.

parallel to stratigraphy, but can also have planar geometry where they cross-cut stratal reflections.

Apparent folding of the diagenetic fronts (Fig. 3b)

At a regional scale, the diagenetic fronts take the form of a series of antiforms and synforms. These undulations in the diagenetic front are coincident with folds that are clearly of tectonic origin. Synforms in diagenetic fronts are coincident with synclines in the tectonic folds and antiforms in the fronts are coincident with anticlines in the tectonic folds (Fig. 4a). The ‘folds’ in the fronts have an amplitude of 20–50 m versus the amplitude of the folded strata of 50–150 m. The amplitude of the folding in the diagenetic front is generally 35% that of the tectonic fold. On seismic lines where synformal and antiformal undulations in the diagenetic fronts occur the fronts are also parallel to the Late Miocene Unconformity.

Serrated pattern (Fig. 3c)

These are asymmetrical sawtooth to stair-step patterns that occur only where the diagenetic fronts cross-cut inclined and deformed stratigraphy (Fig. 2a, b). The ‘teeth’ are between 10 and 50 m in height. Their angle is the same as the inclination of the strata that the front cross-cuts. The width of the serrated patterns is equal to the lateral extent of inclined strata that the fronts pass through. Similar features have been described previously by Davies & Cartwright (2002) and Davies (2005).

Wings (Fig. 3d–f)

Wings are concave-upwards to elongate features in cross-sectional view. Two types of wings have been identified: attached and detached wings. They are differentiated by shape, size and their spatial relationship to the main diagenetic front above which they form.

Attached wings (Fig. 3d) are flat to concave-upward forms, with margins or ‘wings’ that point away upward from the main

front. The bases of these wings are attached to the main front. Attached wings are commonly up to *c.* 0.3–0.8 km in length. The concave-upward forms occur only on the opal-A to opal-CT front (Fig. 2a, c). The flat, less well-developed, form can occur on both fronts, but generally lacks the wings.

Detached wings (Fig. 3e and f) are sharp jumps in the diagenetic front that have slight concave-upward and elongate cross-sectional profiles. This type of wing can be composed of up to three progressively smaller stacked reflections, which are termed tiers (Fig. 4). Detached wings are distinct from attached wings as they form regions where the front becomes elevated and detached from the main front. Detached wings are generally larger in size, but not always, and less concave in shape than attached wings (Fig. 5). Single-tiered detached wings (Fig. 5) appear to be larger versions of attached wings that have separated from the main front. Multi-tiered detached wings (Fig. 4b) in this study tend to form in the troughs of synclines (Fig. 4a, 1). The highest tier of a detached wing can show evidence for a central peak, which marks the highest level of advancement of the feature and diagenetic front (Fig. 4b, CP). Detached wings are normally between 0.8 and 2 km in length.

Terraces (Fig. 3g)

These are regions of the diagenetic front that have stair-step cross-sectional profiles. We distinguish terraces from serrated patterns as terraces form on the margins of larger morphological features (Figs 5 and 6). For example, Davies (2005) identified terraces that formed on the margins of elevated regions of the front that formed a ridge network. We apply the term in the same way here. Terraces are generally 0.25–1 km in length. Wings and terraces are phenomena that generally occur at scales of ≤ 2 km.

Multi-kilometre-scale depressions (Fig. 3h)

Depressions in the diagenetic front that are several kilometres wide have not been described before. The diagenetic front forms a trough-like feature (Fig. 5, 1) that has terraced margins (Fig. 5,

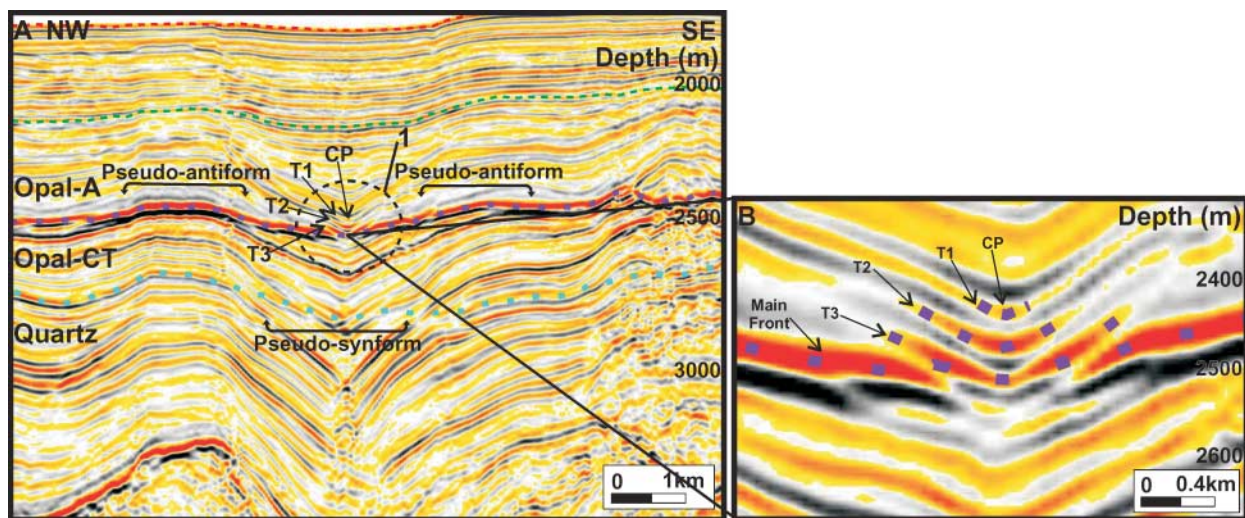


Fig. 4. (a) Seismic line showing apparent folding of the front and an example of a multi-tiered detached wing that has formed in the trough of a syncline (1) that has been cross-cut by the opal-A to opal-CT boundary. (Note that the Late Miocene Unconformity (green dashed line) ‘drapes’ the deformed Miocene strata and that the apparent folding of the diagenetic fronts tracks the unconformity.) (b) Close-up of the multi-tiered detached wing. T1, tier 1; T2, tier 2; T3, tier 3; CP, central peak.

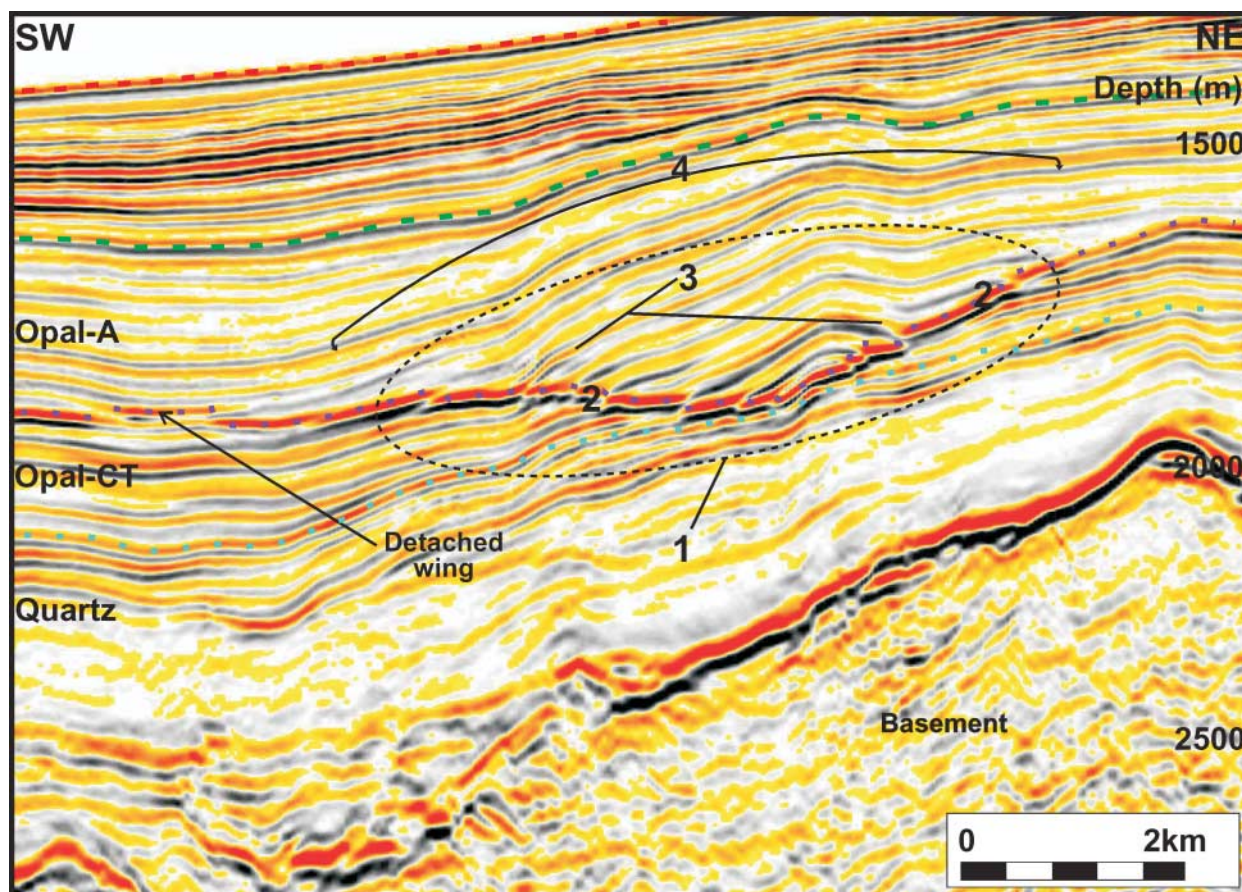


Fig. 5. Seismic line showing an example of a multi-kilometre-scale depression in the opal-A to opal-CT front (1). The depression has terraced margins (2), which can form discontinuous, high-amplitude reflections that extend from the main diagenetic front (3). (Note that the strata above the depression form an antiformal fold (4), and that the opal-CT to quartz boundary has almost merged with the opal-A to opal-CT boundary.) The opal-A to opal-CT boundary also shows an example of a single-tiered detached wing.

2) extending into the depression (Fig. 5, 3). The depressions are 3–5 km in width and 150–250 m in depth. In many cases the depression in the opal-A to opal-CT diagenetic front is deep enough for it to almost merge with the opal-CT to quartz front, which shows no evidence of downward deflection. Where depressions are identified the stratigraphy above the depressed diagenetic front shows evidence for a gentle antiformal fold (Fig. 5, 4). No depressions have so far been identified along the opal-CT to quartz boundary.

Multi-kilometre-scale elevations (Fig. 3i)

Elevations in the diagenetic front are multi-kilometre-scale regions of the front that have preferentially advanced upwards relative to the surrounding regions of the front (Fig. 6, 1). They form arch shapes that cross-cut stratigraphy, hence forming an area of converted sediment in younger stratigraphy. Stratal reflections extend from the outer margins of the elevations producing terraces (Fig. 6, 2). The examples that have been identified are *c.* 4–5 km wide and 50–150 m in height. Where elevations are identified, the stratigraphy above the elevated diagenetic front shows evidence for a gentle synformal fold (Fig. 6, 3). The opal-CT to quartz boundary does not show the same upward deflection.

Evolution of front morphology

Seismic tuning is occurring at reflection discordances but it cannot account for the geometries of the various morphologies observed. This is because the features identified are larger in size than the resolution of the data, which is *c.* 10 m, hence they are not artefacts created by the acquisition and processing of the data. Although seismic tuning will play a role in the seismic amplitudes observed in regions where wings, terraces and serrated patterns form, these features are considered to be real and not seismic artefacts.

The distribution of the frontal morphologies so that they are concentrated in the eastern margin of the basin is probably due to the influence of detrital mineral content. This part of the basin is more distal from the possible detrital source areas (*i.e.* onshore Sakhalin). This means that the diatom ooze would form a greater component of the sediment in the distal areas of the basin, leading to the sediment having a higher silica content. Sediment with a high silica content converts earlier than sediment with an increased detrital component (Isaacs 1981), but silica-rich sediment might also allow greater potential for the various morphologies to develop. Another explanation is that the diagenetic fronts and their associated morphologies are better imaged in the eastern part of the basin.

It is apparent from the examples of the various frontal

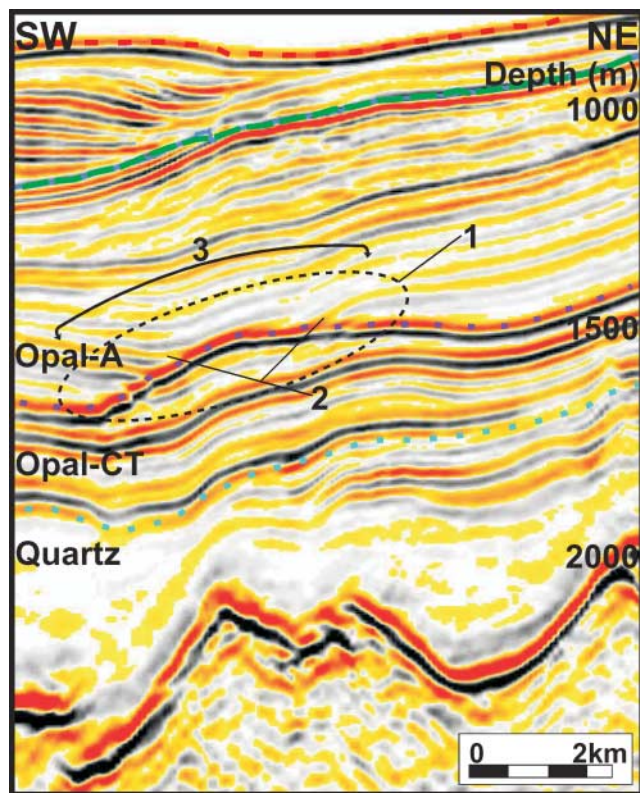


Fig. 6. Seismic line showing an example of a multi-kilometre-scale elevation (1) in the opal-A to opal-CT front. The elevation has terraced margins (2) that extend from the main diagenetic front. (Note the downward flexure of the sediments above the elevation (3).)

morphologies that the opal-A to opal-CT front behaves differently from the opal-CT to quartz front. Most of the geometries are poorly developed and less common along the opal-CT to quartz front, and multi-kilometre-scale depressions and elevations have not been observed along this boundary. This is probably due to the smaller magnitude of the physical property changes, especially the porosity drop, associated with the opal-CT to quartz transition. This could lead to the development of features that are only just above or even below the seismic resolution of the data, and hence are not as obvious on the seismic data.

In the following interpretation it is important to bear in mind that seismically imaged patterns are a snapshot of front advancement, with examples revealing the likely time dependence of diagenetic front cross-sectional development. Diagenetic front patterns should be thought of as dynamic phenomena that evolve over time periods of thousands to millions of years (Tada 1991). We start by briefly considering the apparent folding of the fronts and the parallel relationship with the Late Miocene Unconformity before dealing with each morphology in turn.

Apparent folding of the diagenetic fronts

There are three possible explanations for why the fronts appear to be folded (Fig. 7). For example, the diagenetic fronts could have advanced to their present position after the folding (Fig. 7a, 1–4). Because the Late Miocene Unconformity formed above the syncompressional sedimentary fill of the folds, it forms an

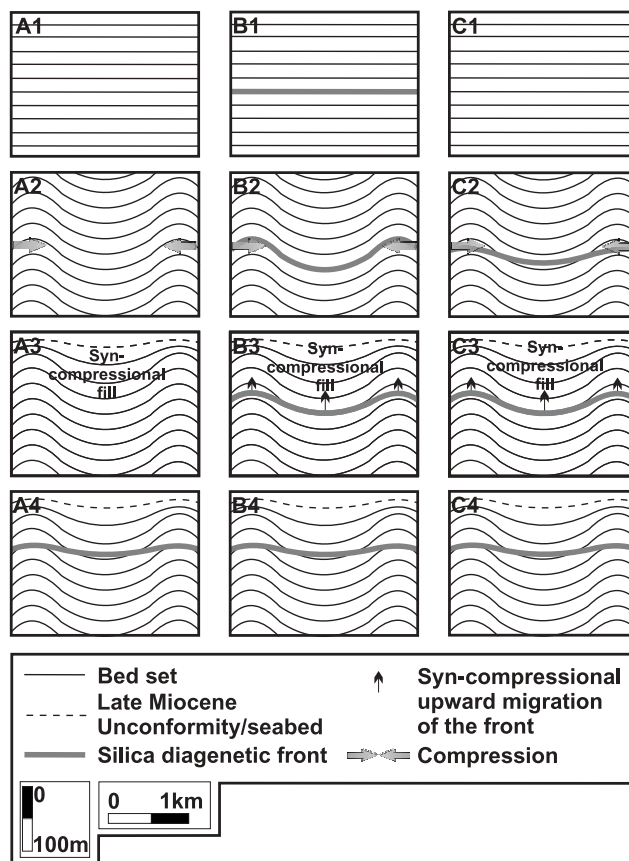


Fig. 7. Schematic diagram showing the possible explanations for the apparent folding of the diagenetic fronts with respect to their parallel relationship to the Late Miocene Unconformity.

undulating surface with the same wavelength as the tectonic folds but a lower fold amplitude (Fig. 7a, 3). As the reactions are principally controlled by temperature the fronts should be parallel to isotherms and these will often be parallel to the contemporaneous sea bed (the Late Miocene Unconformity) (Fig. 7a, 4). Therefore the fronts may simply develop the same undulating relief as the Late Miocene Unconformity. Alternatively, the diagenetic fronts could form before the folding (Fig. 7b, 1–4) and during an episode of lateral compression both the strata and the fronts are deformed (Fig. 7b, 2). Syncompressional sediments are then deposited, mantling the folds. Again, the diagenetic fronts will probably advance to develop a parallel relationship with the contemporaneous sea bed and thereby develop an undulating morphology (Fig. 7b, 3) that tracks the Late Miocene Unconformity (Fig. 7b, 4). Lastly, the fronts may have advanced near the end of the shortening phase (Fig. 7c, 2) and therefore underwent less shortening (Fig. 7c, 3). Once the fronts have regained equilibrium, with the prevailing thermal and chemical conditions, they will track an isotherm parallel to the Late Miocene Unconformity (Fig. 7c, 4).

Our preferred explanation for the occurrence of apparent folding along the diagenetic fronts is that it is probably a result of the parallel relationship with the Late Miocene Unconformity. This specific stratigraphic relationship indicates that the unconformity represented the sea bed, and that rather than being folded the fronts simply track the undulating Late Miocene sea bed. Hence, during the Late Miocene the diagenetic fronts advanced

maintaining a parallelism with the contemporaneous sea bed. Davies & Cartwright (2002) proposed a very similar explanation for the apparent folding of silica diagenetic fronts in the Faeroe–Shetland Basin.

Parallel relationship to the Late Miocene Unconformity

The parallel relationship to the unconformity shows that the depth of the boundaries below the sea bed was constant during the Late Miocene and Early Pliocene. This suggests that they were following isotherms at this time. The reason why the boundaries have not advanced so as to become parallel to the present-day sea bed is uncertain, but it is a phenomenon common to other basins, such as the Faeroe–Shetland Basin (Davies & Cartwright 2002).

Planar fronts

A planar front generally occurs where the diagenetic fronts track stratal boundaries and has not developed any of the morphological features. One could assume that the conversion is therefore occurring at the same time along the beds and bedsets perhaps as a result of similar chemistry within beds and bedsets. However, in reality it is very unlikely that this is true. It is more probable that the reaction occurs in some parts of a bedset first, with the front advancing laterally parallel to bedding until the bedsets are converted. In other words, some sort of differential vertical advancement of the diagenetic front is likely to occur, followed by lateral advancement of the front. This progressive lateral advancement along bedsets would lead to diagenetic fronts being parallel to stratal reflections and would, therefore, inherit the stratigraphic architecture (Fig. 8). This is consistent with some of

the known controls on conversion (sediment age, detrital mineral content, and the host rock chemistry), which locally are likely to be similar within any one bed or bedset, hence leading to lateral advancement, but vary significantly across bedding if the succession is heterogeneous (Davies 2005).

Serrated pattern

Serrated patterns (Fig. 2b and Fig. 8, 1) indicate the preferential advancement of the conversion boundary at a regular separation of a few tens of metres. At this scale the pattern is unlikely to be caused by variations in heat flow. However, it is well known that the detrital mineral content can affect the rate of conversion (e.g. Kastner *et al.* 1977; Isaacs 1982); therefore the simple explanation for the regularity of serrated patterns is variations in the concentration of opal-A within the host sediment. Such variations are probably the result of variations across bedding. The front would produce the serrated pattern by advancing up silica-rich beds faster than silica-poor beds or there being a shallower DOC in richer beds. This indicates that the succession is interbedded and that bedding planes are inclined. If the succession was massive it is likely that serrated patterns would not form.

However, the fact that the teeth of the serrated pattern do not significantly vary in height (Fig. 2a and b), unlike those described by Davies & Cartwright (2002), suggests that this pattern might be an interference effect resulting from the interaction of the reaction occurring at the diagenetic front within inclined and deformed stratigraphy rather than the consequence of detrital mineral concentration. The effect of detrital mineral concentration within the host succession may play only a minor role in the formation of serrated patterns in the North Sakhalin Basin.

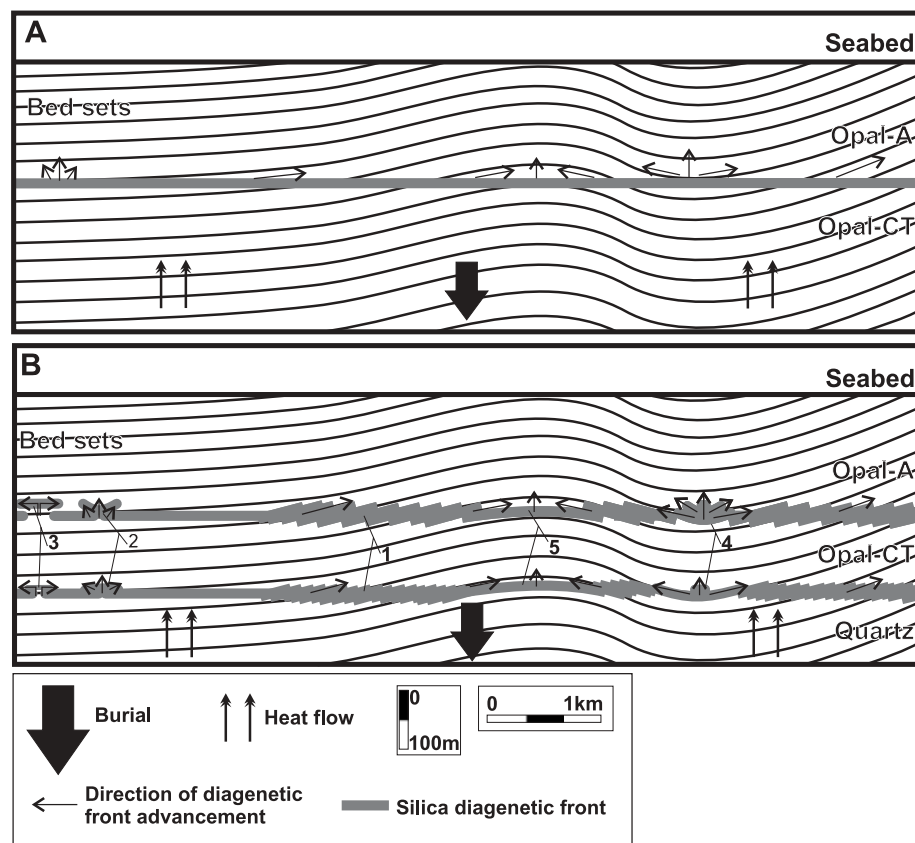


Fig. 8. Schematic diagram showing the mechanisms of silica diagenetic front development in terms of direction of front advancement and inclination of the strata. (a) A combination of temperature and burial cause biogenic silica-rich strata to reach the DOC, where the sediment is transformed from opal-A to opal-CT, forming a diagenetic front. (b) In inclined strata the front advances along silica-rich bedsets, forming serrated patterns (1). In predominantly horizontal strata, attached wings (2) and single-tiered detached wings (3) can form. In folded strata, multi-tiered detached wings can develop in the troughs of synclines (4). However, in the crests of anticlines no detached wings form and the front cross-cuts the fold (5), with serrated patterns forming where the front cuts the limbs of the fold. (Note that the morphological features are less well developed on the opal-CT to quartz diagenetic front.)

Wings

Attached wings (Fig. 8, 2) describe localized concave- and dome-shaped features that form in stratal reflections above the main level of the diagenetic front, but are still attached to the front. The term 'attached' relates to the appearance of the feature on seismic data. They are localized regions where overburden beds and bedsets have started to undergo conversion, but this conversion has not yet advanced laterally along that bed or bedset. The ideal concave shape of an attached wing probably forms as a result of differential subsidence of the central part of the wing, which has undergone conversion before the edges. The tapering edges of a concave-shaped wing probably represent the leading edge of the reaction front.

Detached wings form for exactly the same reasons as attached wings, but we prefer to use this term to indicate that the feature has completely detached from the diagenetic front. The term detached refers to the spatial relationship, in seismic data, to the main diagenetic front that the wing has formed above. It is probable that single-tiered detached wings (Fig. 8, 3) are more evolved forms of attached wings, which signify that the conversion has started to advance laterally along a bed or bedset above the main diagenetic front. As wings advance laterally and convert more sediment one would expect them to flatten out as a result of progressive differential compaction and subsidence. The dominance of lateral advancement and progressive differential subsidence means that the detached wing does not exhibit the concave shape of an attached wing, but still does show the tapering edges that represent the leading edge of the reaction.

A multi-tiered detached wing can form in the trough of a syncline (Fig. 4 and Fig. 8, 4). The front advances through multiple beds and bedsets that have been folded below the DOC, leading to the conversion of several bedsets. This occurs progressively as the front advances both laterally along the limbs and directly upwards from the hinge zone, perhaps as a result of fracture generation causing enhanced diagenesis in the trough of the syncline (see 'Role of fracturing and brecciation' below). The conversion process of the bedsets in the trough of the syncline is time-dependent and would lead to a set of increasingly smaller detached wings stacking one on top of another, with the central peak indicating the highest and most recent stratal level the front has reached (hence the least converted bedset). The concave shape is similar to that of attached wings, but is probably due to a combination of the front following the inclination of the bedsets in the trough of the syncline and some differential subsidence at the centres of the individual tiers.

Terraces

Terraces form in conjunction with kilometre-scale depressions and elevations in the diagenetic front; hence these are explained below along with the larger features.

Depressions and elevations in the diagenetic front

The potential mechanisms for causing depressions and elevations are dealt with together. The multi-kilometre-scale depressions (Fig. 9, 1) in the opal-A to opal-CT diagenetic front have to be consistent with the absence of a depression in the opal-CT to quartz boundary. This would probably rule out lateral variations in heat flow, as we would expect to see a depression in the opal-CT to quartz boundary. Hence, the formation of depressions in the diagenetic front is primarily due to the direction and rate of diagenetic front advancement, with the inclination of the strata

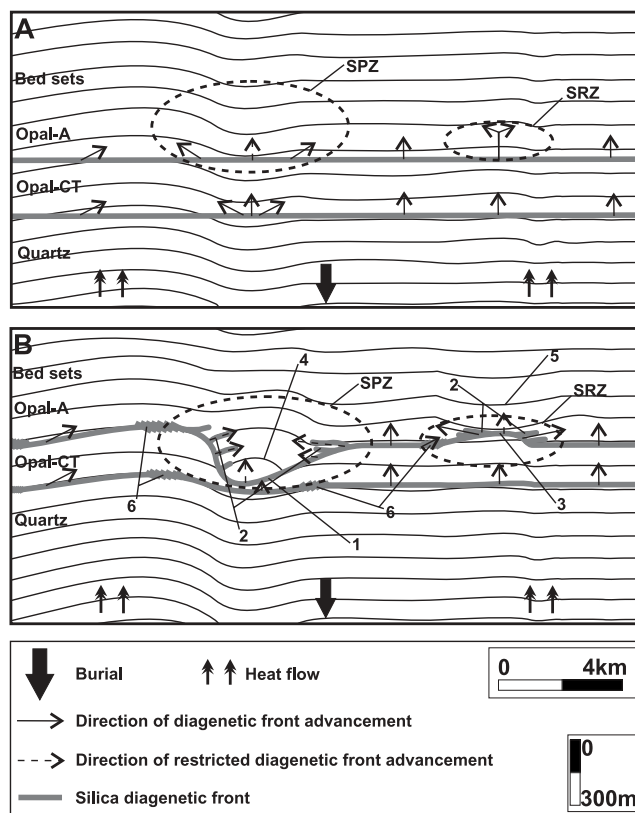


Fig. 9. Schematic diagram showing how depressions and elevations in the diagenetic front form. (a) Diagenetic fronts advance relatively upward through strata that contain zones of silica-poor (SPZ) and silica-rich (SRZ) sediment. (b) Formation of depressions (1) and elevations (3) in the diagenetic front. Terraces (2) form in the unconverted sediment on the margins of depressions and elevations. Differential compaction also occurs in the sediments above these features: antiformal folds (4) above the depressions and downward flexures (5) above the elevations. In inclined strata, serrated patterns form (6).

having little or no influence (Fig. 9). There are two possible hypotheses for these features. First, depressions could form as a result of variations in host lithology geochemistry (e.g. silica, detrital and carbonate content and pore water chemistry). For example, there may be regions that have low concentrations of biogenic silica (Fig. 9, SPZ). This would slow the rate of front advancement relative to adjacent successions. This impeded section of the front forms the depression in the front, as the rate of front advancement outside the silica-poor zone is not impeded.

It is also plausible that the rheology of the sediment is different; it may not be conducive for the development of hydrofractures, which may be important for enhancing the rates of diagenesis (see Eichhubl & Boles 1998). Therefore, a depression may form where the rate of front advancement is inhibited by reduced hydrofracture formation. Gross (1995) suggested that the fracture style within the Monterey Formation, California, depended upon the mineralogical composition of the host lithology. For example, opening-mode veins tended to form in more brittle units such as diagenetic opal-CT beds, whereas faults were more common in beds rich in opal-A and 'weak' detrital minerals (e.g. clays). Hence the mechanical properties of the host lithology can influence the type and extent of fracture formation.

The depression caused by the restriction of the opal-A to opal-CT front can lead to the opal-CT to quartz front that has not been impeded almost merging with the opal-A to opal-CT front. Terraces (Fig. 9, 2), form within the trough of the depression as the front attempts to regain equilibrium with the prevailing burial and geothermal conditions by trying to advance laterally through the fracture- or silica-poor sediment and close the depression to form a planar front.

In contrast to depressions, elevations in the opal-A to opal-CT diagenetic front (Fig. 9, 3) could be accounted for by the front advancing through areas that are particularly rich in biogenic silica (Fig. 9, SPZ). Or, the region may have rheological characteristics that make it prone to hydrofracturing, allowing increased rates of diagenesis. The lack of an elevation in the opal-CT to quartz front rules out lateral variations in heat flow as a possible cause of multi-kilometre-scale elevations. This is despite the fact that the example of an elevation shown in Figure 6 is situated above a basement high, which could suggest that lateral heat flow variations might be responsible for the development of elevations. However, we believe that this spatial arrangement between the opal-A to opal-CT front and basement high is coincidental. This is because of the lack of a spatially coincident elevation in the opal-CT to quartz front, the lack of basement highs underlying other examples of elevations (Fig. 6 is the clearest example of a multi-kilometre-scale elevation), and because nearby examples of depressions in the opal-A to opal-CT front are not underlain by basement lows.

Variations in heat flow may be responsible if there was a difference in the reaction kinetics of the opal-A to opal-CT and opal-CT to quartz transitions. For example, the kinetics of the opal-A to opal-CT transition may mean that the rate of conversion is affected more by heat flow variations than the opal-CT to quartz transition, so that the transition could be slower (forming a depression) or faster (forming an elevation) than the less affected opal-CT to quartz transition. This scenario is highly unlikely, as there is no evidence in the literature for a difference in the reaction kinetics, with only host lithology, such as detrital mineral content (e.g. Isaacs 1982), and pore water chemistry (i.e. sedimentological factors) leading to variations in the conversion rate.

The elevation would form where the front has initially been able to advance more rapidly upwards than the surrounding sections of the front, forming a multi-kilometre-scale arch shape. The advancement at the elevation has subsequently been restricted, by mechanical–chemical factors (such as silica content or fracture formation), causing the front to preferentially advance laterally along beds and bedsets, forming terraces along the margins of the elevation (Fig. 9, 2). The lateral advancement of terraces extending from the elevation along adjacent unconverted bedsets would allow the rest of the front to advance and reach the same stratigraphic level as the elevation by converting the sediment above the main front. Once the front passes the zone of favourable lithology, forming the elevation, the front reaches equilibrium with its surrounding conditions (i.e. host lithology, temperature and burial rate) and no longer undergoes differential advancement. Elevations and depressions are examples of where the lithological controls occur over a large area.

Front advancement processes

The above features provide a record of diagenetic front advancement within a clastic sedimentary basin. Together with outcrop analogues and siliceous sediment exposed elsewhere (e.g. Monterey Formation, California) they can provide clues as to what

processes and interrelated processes are important in front advancement.

Bedding and lithology control

A variety of outcrop and borehole data show that diatomites are commonly laminated and interbedded at centimetre to metre scale (e.g. Compton 1991; Tada 1991). Outcrop evidence from the Monterey Formation also indicates that brecciation and microfracturing can accompany the reaction as a result of hydrofracturing and volumetric contraction (Eichhubl & Boles 1998; Eichhubl 2004). Sediments richer in silica are converted at a faster rate than those containing a greater proportion of detrital minerals (e.g. Isaacs 1981, 1982; Bohrmann *et al.* 1994). It is reasonable to propose that many patterns probably form as a result of bedding and lithology variations, with conversion occurring in beds and bedsets that are richer in opal-A first, either at a slightly shallower DOC or as a result of a higher rate of front advancement, than in beds that have a higher detrital mineral content. This simple mechanism has already been shown as one of the ways to account for the serrated patterns of the fronts in inclined strata.

In contrast, wings form where the diagenetic front has ‘jumped’ to a shallower stratigraphic level. We term these localized regions where the reaction has been initiated in overburden strata ‘seed areas’. There are several mechanisms that could account for the development of seed areas (Fig. 10).

The silica diagenetic transformations are dehydration reactions that can generate excess pore pressure in the strata above the diagenetic fronts (Hesse 1990). This generation of overpressure could cause hydrofracturing in the sediments overlying the fronts. Opening of fractures by hydrofracturing of materials such as diatomites with very low permeability occurs when the fluid pressure is so high that the minimum principal effective stress, σ'_3 , has been reduced to zero and the tensile stresses exceed the cohesive strength of the sediment (Bjørlykke & Høeg 1997). Seed areas may form where the density of hydrofracture generation is particularly high (Fig. 10a), causing enhanced rates of diagenesis (Eichhubl & Behl 1998) leading to localized differential front advancement. Zones of denser hydrofracture generation may result from lateral variations in sediment rheology or differences in pore pressure in the overlying strata.

The concentration of opal-A in overburden sediments may vary and seed areas would preferentially develop in areas where the silica concentration is higher (Fig. 10b), leading to differential front advancement. It could also be that part of the bedset that the front has reached is thinner than the adjacent areas (Fig. 10c). Because there is less sediment for the front to convert at the thinnest points of a bedset, the front would preferentially advance to the next bedset at these areas. It is likely that all of the mechanisms that can generate seed areas play a role in the formation of wings, but without well calibration of the succession the exact mechanisms and their relative importance remain unclear.

Role of fracturing and brecciation

It is well known that fracturing enhances the rate of the diagenetic process (Eichhubl & Behl 1998). Microfractures can form because of the volumetric contraction as a result of the conversion of opal-CT to quartz (Eichhubl 2004). Fractures also form as a result of overpressure created by the expulsion of fluids by the reaction (Hesse 1990), which causes hydrofractures that are beyond seismic resolution to propagate in the overlying

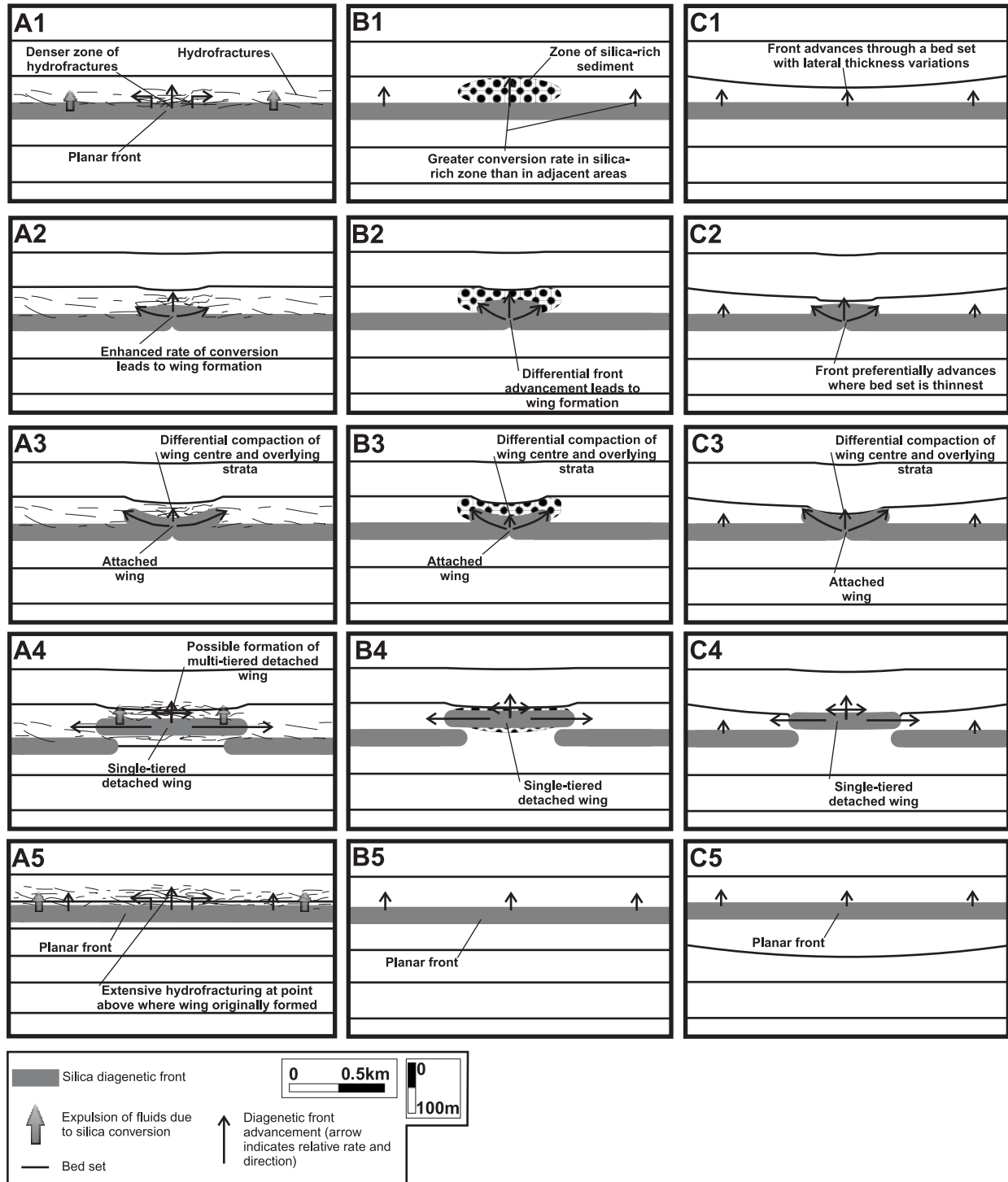


Fig. 10. Schematic diagram showing the mechanisms of seed area development. The diagenetic front initially tracks the stratigraphy. (a) Seed area generation as a result of increased hydrofracture generation. (b) Seed area generation as a result of a zone of silica-rich sediment. (c) Seed area generation as a result of bedset thickness variations. Sequences 1–5 show the formation of attached wings by differential advancement at a seed area and subsequent evolution into single-tiered detached wings and then via lateral advancement, and joining with other wings at that stratal level, into planar fronts.

sediment (see ‘Bedding and lithology control’ above). Comparisons with outcrop examples, especially the Monterey Formation, indicate that hydrofracturing varies as a function of lithology. The increase in pore fluid pressure caused by the reactions would provide the conditions for brittle fracture, but as a result of the effect that detrital content has on conversion the lithologies

would fracture at different times and depths (Eichhubl & Behl 1998). The rheology of the lithology would have an effect on the type and extent of fracturing (e.g. Gross 1995); for example, well-cemented and competent lithologies are more susceptible to fracturing than relatively weak lithologies such as mudstones.

Hydrofractures in diatomites stimulated by water injection

(Barenblatt *et al.* 2002) are predominantly parallel to bedding; they provide natural planes of weakness for mechanical failure, and cause regions of significant brecciation and considerably enhanced permeability. This would also provide a mechanism for the advancement of the diagenetic front parallel to bedding. Therefore, hydrofractures may be important in the development of the geometries described in this paper. The orientation of naturally occurring hydrofractures is generally normal to the direction of σ'_3 unless the sediment is very anisotropic (Bjørlykke & Høeg 1997). These networks of hydrofractures can occur along individual beds or at several different levels and are termed 'damage zones' (Barenblatt *et al.* 2002). It is known that increased permeability can lead to increased fluid flow, which in turn can cause enhanced dissolution and reprecipitation of silica (Williams *et al.* 1985), and hence enhanced rates of conversion. Similar brecciated damage zones, triggered by the liberation of water and overpressure, are also seen in outcrops of the siliceous Monterey Formation, where they caused enhanced rates of diagenesis (Eichhubl & Behl 1998; Eichhubl & Boles 1998).

Hydrofracturing caused by silica diagenesis can influence where subsequent conversion can occur, and hence the points along the diagenetic front at which certain morphologies can form. Therefore, the development of localized denser fracture networks may provide a mechanism for the growth of seed areas, the formation of attached wings and subsequent development into detached wings and finally into wider regions of planar fronts that track stratigraphy (Fig. 10a, 1–5). Once the original wing has developed into a planar front there may be a relict denser hydrofracture network above the area where the seed area and wing originally formed (Fig. 10a, 5); this area would provide an optimum location for subsequent seed area generation. Hence, front advancement and wing formation as a result of hydrofracture generation could result in the development of a feedback mechanism.

Deformation caused by differential front advancement

Davies (2005) showed that differential advancement of the opal-A to opal-CT diagenetic front can cause folding in the overburden. This is because of the significant porosity loss that the reaction causes, such that areas of preferential front advancement are overlain by depressions caused by differential subsidence (Fig. 11). However, above many of the smaller-scale morphological features in the North Sakhalin Basin this effect is not observed, probably for two reasons: either the magnitude of the porosity change could be so small that the differential compaction effect described by Davies (2005) has no observable effect (i.e. it is below the seismic resolution of the dataset), or, at geological time scales, the reaction is occurring at essentially the same time along the length of the beds and bedsets. In many places the front tracks the stratigraphy; hence the deformation of the overlying sediments can be considered to be transient deformation, which leads to temporary folding as the diagenetic fronts convert entire bedsets, removing the depressions formed by areas of preferential front advancement (Fig. 11).

However, kilometre-scale differential compaction in the North Sakhalin Basin has been observed where depressions and elevations in the opal-A to opal-CT diagenetic front develop. The subtle antiforms occurring above the depressions (Fig. 5, 4, and Fig. 9, 4) and the downward flexures above the elevations (Fig. 6, 3, and Fig. 9, 5) are both interpreted to be the result of differential front advancement and therefore differential compaction.

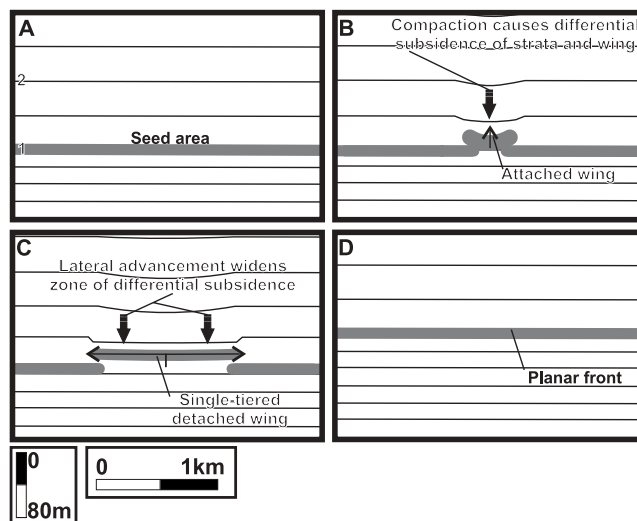


Fig. 11. Schematic diagram showing how differential front advancement can cause transient deformation in the overlying sediments. The diagenetic front (1) tracks the stratigraphy (2). It should be noted that the bedsets below the diagenetic front are more compacted than those above, as a result of the change in physical properties.

Discussion

3D geometry

As we use 2D seismic data we have not made a judgement on the likely plan-form geometry of the features. It is likely that the terraces and serrated patterns probably form ridges in plan form that are parallel to the strike of the folds. This is based upon the occurrence of similar phenomena mapped in three dimensions by Davies (2005) in the Faeroe–Shetland Basin. The development of a comprehensive 2D and 3D scheme would be useful for future descriptions of silica diagenetic fronts.

Controls on diagenetic front morphology and comparison with the Faeroe–Shetland Basin

The silica diagenetic boundaries in Sakhalin are amongst the largest diagenetic front patterns to be described. Many of the morphological features appear to form as a result of the interference of the front with a layered stratigraphy, rather than taking up the cell-like or polygonal geometry described earlier in the basins of the NE Atlantic margin (Davies 2005).

The morphologies in the Faeroe–Shetland Basin take the form of a series of polygonal ridges composed of attached wings and terraces with polygonal or cell-like plan-form geometry. Although the morphologies of the North Sakhalin Basin include attached wings and terraces, there is no evidence for the development of polygonal ridges with the associated 3D planar form. Both basins have a similar heterogeneous stratigraphic fill, which is composed of diatomites interbedded with diatomaceous mudstones. The morphologies of the two basins are of similar scale, generally being of the order of <2 km, with the exception of the multi-kilometre-scale depressions and elevations of the North Sakhalin Basin. However, the distribution of the morphologies for the basins is different. In the North Sakhalin Basin the morphologies develop throughout the basin wherever the silica diagenetic boundaries occur, although they are more commonly developed in the eastern part of the basin, whereas in the Faeroe–Shetland Basin the polygonal ridges are developed only

in certain areas even though the opal-A to opal-CT transition occurs over a much larger extent of the basin. Davies (2005) noted that in the Faeroe–Shetland Basin the siliceous sediments in which the boundaries formed the cell-like geometry mantled an underlying polygonal fault system and that ridges in the diagenetic front generally formed above the fault tips, probably as a result of fluid flow from the faults causing enhanced rates of diagenesis. However, in the North Sakhalin Basin there is no polygonal fault system that can influence formation of any of the morphological features; hence this specific ridge morphology did not develop.

Therefore, it is likely that the morphologies of the North Sakhalin Basin are a more fundamental manifestation of the processes of silica diagenetic front advancement than the more specialized process derived from underlying cellular fracture patterns by Davies (2005) in the Faeroe–Shetland Basin. Differential front advancement is more common in the Faeroe–Shetland Basin because of the regularity of morphology development associated with the extensive polygonal faulting. The irregular nature of morphology development in the North Sakhalin Basin means that the morphologies are not as commonly developed as those in the Faeroe–Shetland Basin; hence the fronts tend to mostly follow bedsets when the strata are planar, with differential front advancement occurring only where seed areas form.

It is probable that the positions of seed area formation in the North Sakhalin Basin are controlled by a complex natural system, which is influenced by a combination of factors. There may be one dominant factor governing the formation and position of the various morphological features, or a combination of factors, but it is impossible to tell using seismic data alone.

Remaining questions

The parallel relationship of the diagenetic fronts to the Late Miocene Unconformity suggests that the fronts have not advanced significantly since the Late Miocene. The mechanisms responsible for the development of the diagenetic fronts, including the reasons for the parallel relationship, and their present-day sluggish behaviour are unclear. Davies & Cartwright (2002) suggested that there was an elevated palaeogeothermal gradient in the Faeroe–Shetland Basin when the conversion was initiated and that the geothermal gradient is lower at the present day, leading to the front appearing sluggish and being parallel to an Early Pliocene unconformity. However, Davies & Cartwright (2002) were wrong in their hypothesis that the diagenetic front was fossilized and no longer advancing; the reaction would not stop, but instead there has been a significant decrease in the rate of conversion, which produces the apparent sluggish behaviour. No palaeogeothermal data currently exist for the North Sakhalin Basin, hence a higher geothermal gradient cannot be ruled in or out as a cause for the apparent sluggish behaviour. There may be several other causes for the apparent sluggish behaviour such as (1) erosion of overburden (producing a similar effect to a temperature decrease), (2) an increase in sedimentation rates relative to the rate of conversion, or (3) a change in the rate of conversion, and hence the rate the front advances, as a result of variations in physio-chemical factors (such as the opal-A content of the sediment). Further investigation of these issues, including well calibration, is required to solve this problem.

Further detailed borehole data are required to assess whether our hypothesis, that diagenetic front geometry at regional scales is controlled by a combination of lithological variation and the interference of the fronts with deformed and inclined strata, is

correct. Borehole data would also help to evaluate the role of fracturing in front advancement. Scientific drilling of some of the phenomena described would be very useful.

Conclusions

Silica diagenetic fronts are dynamic, time-dependent phenomena that have mostly been overlooked in terms of their importance in the evolution of sedimentary basin fills. Seismic analysis has a significant role to play in understanding how the fronts advance and how this advancement changes the properties and structure of the host sediment.

The silica diagenetic boundaries of the North Sakhalin Basin reveal a number of morphological features that form as a result of the interference of the diagenetic front with inclined and folded stratigraphy as well as with variations in composition and rheology of the host lithology. The morphological elements can be explained in terms of variations in the relative rates of front advancement, both parallel and orthogonal to bedding, which are probably controlled by lithological variations, the inclination of the host strata, and feedback between diagenesis and possible hydrofracture generation. Wings, terraces and serrated patterns are all indicative of the attempts by the fronts to advance to higher stratal levels, whereas multi-kilometre-scale depressions and elevations in the opal-A to opal-CT front probably result from the rate of front advancement being restricted or enhanced by localized variations in host lithology.

The fronts of the North Sakhalin Basin lack the regular cell-like or polygonal geometry of the basins of the NE Atlantic margin (e.g. Davies 2005), as a result of different controls on seed area formation and front advancement. This has led to an irregular geometry of the diagenetic fronts. These varied front patterns between basins indicate that the factors affecting front advancement and the growth of the various morphological features differ depending on the stratigraphic, sedimentological and structural architecture of a particular basin. Therefore, it can be concluded that within individual sedimentary basins there could be unique frontal geometries as a result of the variations in the controls on the diagenetic fronts.

The factors that influence the geometry of the fronts include the location and extent of fluid flow conduits (e.g. hydrofractures), pore water chemistry, host rock lithology and variations in heat flow. Within the North Sakhalin Basin the interaction of all these factors, together with the lack of one dominant control, during the diagenetic transitions created the irregular frontal geometry that is observed. The precise diagenetic processes cannot be accurately determined using seismic data alone. However, the visualization of diagenetic fronts within a basin using seismic reflection data is very useful at this scale.

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References

- BARENBLATT, G.I., PATZEK, T.W., PROSTOKISHIN, V.M. & SLIN, D.B. 2002. Oil deposits in diatomites: A new challenge for subterranean mechanics. *In: SPE/DOE 13th Symposium on Improved Oil Recovery 13–17 April 2002 Society*

- of Petroleum Engineers, **75230**.
- BJØRLYKKE, K. & HJØEG, K. 1997. Effects of burial diagenesis on stresses, compaction and fluid flow in sedimentary basins. *Marine and Petroleum Geology*, **14**, 267–276.
- BOHRMANN, G., ABELMANN, A., GERSONDE, R., HUBBERTEN, H. & KUHN, G. 1994. Pure siliceous ooze, a diagenetic environment for early chert formation. *Geology*, **22**, 207–210.
- CHAIKA, C. & DVORKIN, J. 2000. Porosity reduction during diagenesis of diatomaceous rocks. *AAPG Bulletin*, **84**, 1173–1184.
- COMPTON, J.S. 1991. Porosity reduction and burial history of siliceous rocks from the Monterey and Sisquoc Formations, Point Pedernales area, California. *Geological Society of America Bulletin*, **103**, 625–636.
- DAVIES, R.J. 2005. Differential compaction and subsidence in sedimentary basins due to silica diagenesis: A case study. *Geological Society of America Bulletin*, **117**, 1146–1155.
- DAVIES, R.J. & CARTWRIGHT, J. 2002. A fossilized Opal A to Opal C/T transformation on the northeast Atlantic margin: Support for a significantly elevated palaeogeothermal gradient during the Neogene? *Basin Research*, **14**, 467–486.
- DAVIES, R.J. & CLARK, I. 2006. Continental slope failures primed and triggered by silica diagenesis: A case study based upon 3D seismic data analysis. *Basin Research*, **18**, 339–550.
- DAVIES, R.J., HUISE, M., HIRST, P., CARTWRIGHT, J. & YANG, Y. 2006. Giant clastic intrusions primed by silica diagenesis. *Geology*, **34**, 917–920.
- EICHHUBL, P. 2004. Growth of ductile opening-mode fractures in geomaterials. In: COSGROVE, J.W. & ENGELDER, T. (eds) *The Initiation, Propagation, and Arrest of Joints and Other Fractures*. Geological Society, London, Special Publications, **231**, 11–24.
- EICHHUBL, P. & BEHL, R.J. 1998. Diagenesis, deformation, and fluid flow in the Miocene Monterey Formation. In: EICHHUBL, P. (ed.) *Diagenesis, Deformation, and Fluid Flow in the Miocene Monterey Formation*. Pacific Section, SEPM, Special Publications, **83**, 5–13.
- EICHHUBL, P. & BOLES, J.R. 1998. Vein formation in relation to burial diagenesis in the Miocene Monterey Formation, Arroyo Burro Beach, Santa Barbara, California. In: EICHHUBL, P. (ed.) *Diagenesis, Deformation, and Fluid Flow in the Miocene Monterey Formation*. Pacific Section, SEPM, Special Publications, **83**, 15–36.
- GROSS, M.R. 1995. Fracture partitioning: Failure mode as a function of lithology in the Monterey Formation of coastal California. *Geological Society of America Bulletin*, **107**, 779–792.
- GUERIN, G. & GOLDBERG, D. 1996. Acoustic and elastic properties of calcareous sediments across a siliceous diagenetic front on the eastern US continental slope. *Geophysical Research Letters*, **23**, 2697–2700.
- HEIN, J.R., SCHOLL, D.W., BARRON, J.A., JONES, M.G. & MILLER, J. 1978. Diagenesis of late Cenozoic diatomaceous deposits and formation of the bottom simulating reflector in the southern Bering Sea. *Sedimentology*, **25**, 155–181.
- HESSE, R. 1990. Origin of chert: Diagenesis of biogenic siliceous sediments. In: MCILREATH, I.A. & MORROW, D.W. (eds) *Diagenesis*. Geoscience Canada Reprint Series, **4**, 227–252.
- HINMAN, N.W. 1990. Chemical factors influencing the rates and sequences of silica phase transitions: Effects of organic constituents. *Geochimica et Cosmochimica Acta*, **54**, 1563–1574.
- ISAACS, C.M. 1981. Porosity reduction during diagenesis of the Monterey Formation, Santa Barbara coastal area, California. In: GARRISON, R.E. & DOUGLAS, R.G. (eds) *The Monterey Formation and Related Siliceous Rocks of California*. Pacific Section, SEPM, Special Publications, **15**, 257–271.
- ISAACS, C.M. 1982. Influence of rock composition on kinetics of silica phase changes in the Monterey Formation, Santa Barbara area, California. *Geology*, **10**, 304–308.
- KASTNER, M., KEENE, J.B. & GIESKES, J.M. 1977. Diagenesis of siliceous oozes—I. Chemical controls on the rate of opal-A to opal-CT transformation—an experimental study. *Geochimica et Cosmochimica Acta*, **41**, 1041–1051.
- KELLER, M.A. & ISAACS, C.M. 1985. An evaluation of temperature scales for silica diagenesis in diatomaceous sequences including a new approach based on the Miocene Monterey Formation, California. *Geo-Marine Letters*, **5**, 31–35.
- KHVEDCHUK, I. 1993. The petroleum basins of the Sea of Okhotsk (abstract). *AAPG Bulletin*, **77**, 1637.
- KURAMOTO, S., TAMAKI, K., LANGSETH, M.G., NOBES, D.C., TOKUYAMA, H., PISCIOOTTO, K.A. & TAIRA, A. 1992. Can opal-A/opal-CT BSR be an indicator of the thermal structure of the Yamato Basin, Japan Sea? In: TAMAKI, K., SUEHIRO, K., ALLEN, J. ET AL. (eds) *Proceedings of the Ocean Drilling Program, Scientific Results, 127/128*. Ocean Drilling Program, College Station, TX, 1145–1156.
- LANGSETH, M.G. & TAMAKI, K. 1992. Geothermal measurements: Thermal evolution of the Japan Sea basins and sediments. In: TAMAKI, K., SUEHIRO, K., ALLEN, J. ET AL. (eds) *Proceedings of the Ocean Drilling Program, Scientific Results, 127/128*. Ocean Drilling Program, College Station, TX, 1297–1309.
- LEE, G.H., KIM, H., JOU, H. & CHO, H. 2003. Opal-A/opal-CT phase boundary inferred from bottom-simulating reflectors in the southern South Korea Plateau, East Sea (Sea of Japan). *Geophysical Research Letters*, **30**, 2238–2241.
- LINDQUIST, S.J. 2000. *The North Sakhalin Neogene Total Petroleum System of Eastern Russia*. US Geological Survey Open-File Report, **99-50-O**.
- NOBES, D.C., MURRAY, R.W., KURAMOTO, S., PISCIOOTTO, K.A. & HOLLER, P. 1992a. Impact of silica diagenesis on physical property variations. In: TAMAKI, K., SUEHIRO, K., ALLEN, J. ET AL. (eds) *Proceedings of the Ocean Drilling Program, Scientific Results, 127/128*. Ocean Drilling Program, College Station, TX, 3–23.
- NOBES, D.C., LANGSETH, M.G., KURAMOTO, S., HOLLER, P. & HIRATA, N. 1992b. Comparison and correlation of physical-property results from Japan Sea basin and rise sites, legs 127 and 128. In: TAMAKI, K., SUEHIRO, K., ALLEN, J. ET AL. (eds) *Proceedings of the Ocean Drilling Program, Scientific Results, 127/128*. Ocean Drilling Program, College Station, TX, 1275–1296.
- PISCIOOTTO, K.A. 1981. Diagenetic trends in the siliceous facies of the Monterey Shale in the Santa Maria region, California. *Sedimentology*, **28**, 547–571.
- SCOTT, S.K. 1994. *Oscillations, Waves and Chaos in Chemical Kinetics*. Oxford University Press, Oxford.
- TADA, R. 1991. Compaction and cementation in siliceous rocks and their possible effect on bedding enhancement. In: EINSELE, G., RICKEN, W. & SEILACHER, A. (eds) *Cycles and Events in Stratigraphy*. Springer, Berlin, 480–491.
- TULL, S.J. 1997. The diversity of hydrocarbon habitat in Russia. *Petroleum Geoscience*, **3**, 315–325.
- VOLPI, V., CAMERLENGHI, A., HILLENBRAND, C.-D., REBESCO, M. & IVALDI, R. 2003. Effects of biogenic silica on sediment compaction and slope stability on the Pacific margin of the Antarctic Peninsula. *Basin Research*, **15**, 339–363.
- WEAVER, R., ROBERTS, A.P., FLECKER, R. & MACDONALD, D.I.M. 2004. Tertiary geodynamics of Sakhalin (NW Pacific) from anisotropy of magnetic susceptibility fabric and paleomagnetic data. *Tectonophysics*, **379**, 25–42.
- WILLIAMS, L.A. & CRERAR, D.A. 1985. Silica diagenesis: II. General mechanisms. *Journal of Sedimentary Petrology*, **55**, 312–321.
- WILLIAMS, L.A., PARKS, G.A. & CRERAR, D.A. 1985. Silica diagenesis: I. Solubility controls. *Journal of Sedimentary Petrology*, **55**, 301–311.
- WORRALL, D.M., KRUGLYAK, V., KUNST, F. & KUZNETSOV, V. 1996. Tertiary tectonics of the Sea of Okhotsk, Russia: Far-field effects of the India–Eurasia collision. *Tectonics*, **15**, 813–826.