

Growth and demise of Permian biogenic chert along northwest Pangea: evidence for end-Permian collapse of thermohaline circulation

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Abstract

The Permian Chert Event (PCE) was a 30 Ma long episode of unusual chert accumulation along the northwest margin of Pangea, and possibly worldwide. The onset of the PCE occurred at about the Sakmarian–Artinskian boundary in the Sverdrup Basin, Canadian Arctic, where it coincides with a maximum flooding event, the ending of high-frequency/high-amplitude shelf cyclicity, the onset of massive biogenic chert deposition in deep-water distal areas, and a long-term shift from warm- to cool-water carbonate sedimentation in shallow-water proximal areas. A similar and coeval shift is observed from the Barents Sea to the northwestern USA. A landward and southward expansion of silica factories occurred during the Middle and Late Permian at which time warm-water carbonate producers disappeared completely from the northwest margin of Pangea. Biotically impoverished and increasingly narrow cold-water carbonate factories (characterised by non-cemented bioclasts of sponges, bryozoans, echinoderms and brachiopods) were then progressively replaced by silica factories. By Late Permian time, little carbonate sediments accumulated in the Barents Sea and in the Sverdrup Basin, where the deep- to shallow-water sedimentary spectrum was occupied by siliceous sponge spicules. By that time, biogenic silica sedimentation was common throughout the world. Silica factories collapsed in the Late Permian, abruptly bringing the PCE to an end. In northwest Pangea, the end-Permian collapse of the PCE was associated with a major transgression and with a return to much warmer oceanic and continental climatic conditions. Chert deposition resumed in the distal oceanic areas during the early Middle Triassic (Anisian) after a 8–10 Ma interruption (Early Triassic Chert Gap). The conditions necessary for the onset, expansion and zenith of the PCE were provided by the thermohaline circulation of nutrient-rich cold waters along the northwestern and western margin of Pangea, and possibly throughout the world oceans. These conditions provided an efficient transportation mechanism that constantly replenished the supply of silica in the area, created a nutrient- and oxygen-rich environment favouring siliceous biogenic productivity, established cold sea-floor conditions, hindering silica dissolution, while increasing calcium carbonate solubility, and provided conditions adverse to organic and inorganic carbonate production. The northwest margin of Pangea was, for nearly 30 Ma, bathed by cold waters presumably derived from the seasonal melting of northern sea ice, the assumed engine for thermohaline circulation.

This process started near the Sakmarian–Artinskian boundary, intensified throughout Middle and Late Permian time and ceased suddenly in latest Permian time. It led to oceanic conditions much colder than normally expected from the palaeolatitudes, and the influence of cold northerly-derived water was felt as far south southern Nevada. The demise of silica factories was caused by the rapid breakdown of these conditions and the establishment of a much warmer marine environment accompanied by sluggish circulation and perhaps a reduced input of dissolved silica to the ocean. Complete thawing of northern sea ice would have ended thermohaline circulation and led to warm and sluggish oceanic conditions inimical to the production, accumulation and preservation of biogenic silica. © 2002 Elsevier Science B.V. All rights reserved.

Keywords: chert; carbonate; Permian; Triassic; Pangea; cooling; thermohaline

1. Introduction

The end-Permian extinction and associated environmental turnover has received much attention in the literature over the past decade (see summaries by Erwin, 1993, 1994, 1995), an interest fueled by the K–T boundary extinction and the extra-terrestrial bolide hypothesis (Alvarez et al., 1980). Collision with a speeding interplanetary projectile (Xu et al., 1985; Xu and Yan, 1993; Retallack et al., 1998; Becker et al., 2001; Kaiho et al., 2001) is but one of numerous causes that have been invoked to explain the end-Permian devastation. Of particular interest are the purported changes to the global oceanic circulation system leading to, or accompanying, the end-Permian event. Diametrically-opposed views have been expressed in the literature. For instance, it has been suggested that the Permian was a time of increasingly sluggish thermohaline circulation (Keith, 1982; Herring, 1995), leading to increasingly oxygen-deficient, and ultimately anoxic, bottom waters, the rapid and presumably global overturn of which contributed to the extinction event (Kajiwara et al., 1994; Knoll et al., 1996; Suzuki et al., 1998). In contrast, it has been proposed that most of the Permian was a time of well-oxygenated thermohaline circulation (Isozaki, 1994, 1997), the demise of which led to a worldwide anoxic event during the Lopingian with global anoxia spreading from deep to shallow environments and extending well into the Early Triassic (Wignall and Hallam, 1992; Wignall et al., 1998).

The goal of this paper is to present evidence that thermohaline oceanic circulation was vigorous along the margins of Pangea during the Per-

mian yet sluggish during the Early Triassic. This evidence is based on the widespread occurrence of Permian biogenic cherts regionally and the absence of Early Triassic cherts globally. This paper builds and expands from the contributions of Beauchamp (1994), who first documented the Permian cooling trend along the northwest margin of Pangea, and Beauchamp and Desrochers (1997), who described and interpreted the range of warm to cold Permian carbonate and chert facies associated with the cooling trend. Both of these contributions suggested that oceanic circulation may have played a major role in the observed cooling, an idea also examined recently by Mei and Henderson (2001) to explain the distribution of conodont fauna. This theme is further developed here.

2. Permian Chert Event (PCE) and Early Triassic Chert Gap (ETCG)

Some episodes of Earth's history have been more favourable to chert formation than others (Hein and Parrish, 1987; Maliva et al., 1990). The term 'burb', as in the 'Eocene Silica Burb' (McGowran, 1989), or the term 'event', as in the 'Permian Chert Event' (Murchey and Jones, 1992), have been used to label some of these episodes when biogenic silica appears to have been produced faster and/or preserved more efficiently than during other periods. Likewise, there were times when very little, if any, biogenic chert was produced and/or preserved. The 8–10 Ma long Early Triassic 'Chert Gap' is one of such events (Casey, 1993; Kakuwa, 1996; Kozur, 1998a,b; Racki, 1999).

2.1. Prior to PCE: Sakmarian and earlier

For most of the Carboniferous period, and up until the Sakmarian (mid Early Permian), biogenic silica factories existed along the northwest margin of Pangea, occurring in distal, deep-water, slope to basinal areas adjacent to productive car-

bonate factories. While biogenic silica production was high during this interval, spiculitic and radiolarian deep-water deposits are not necessarily preserved as chert, but rather as a vast array of cherty carbonates and mudrocks. Examples of such deposits include: the Kuna and Tupik formations and the Akmalik and Imnaitchiak forma-

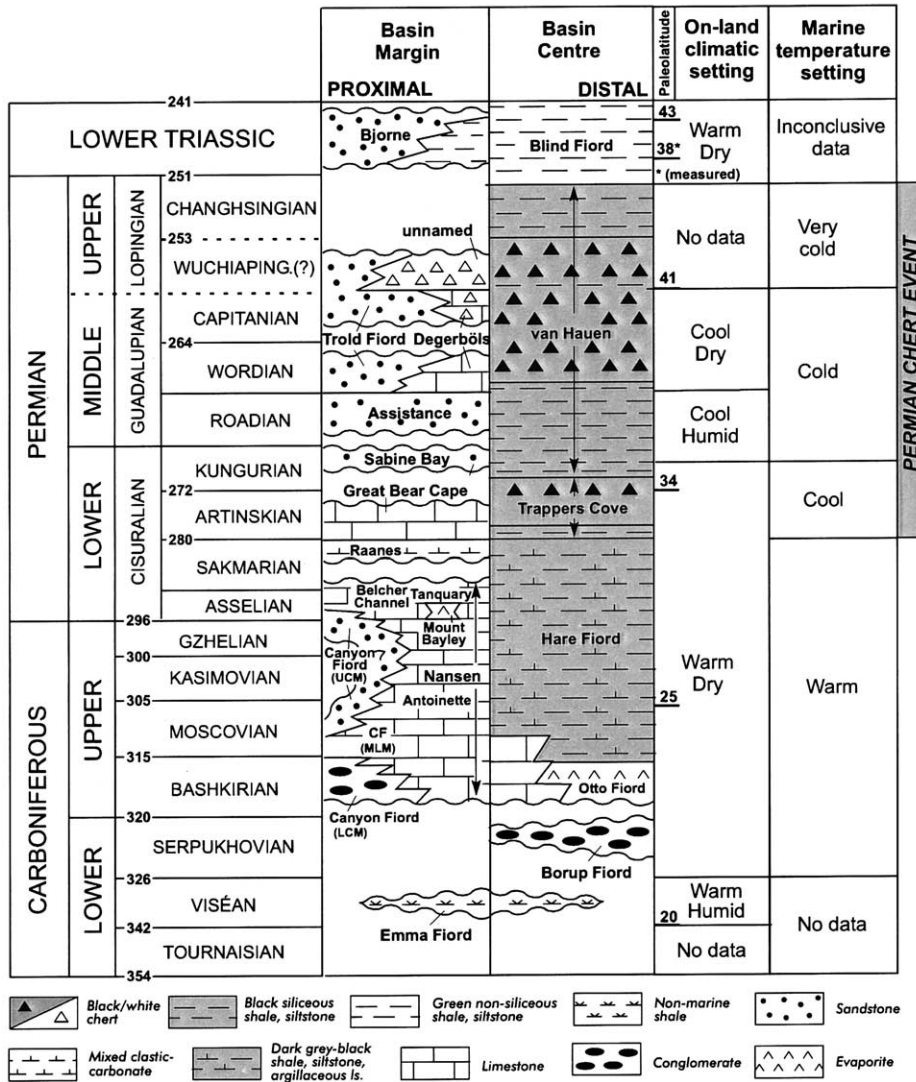


Fig. 1. Stratigraphic chart of Sverdrup Basin showing range of Permian Chert Event against Carboniferous and Permian stages and corresponding radiometric dates (Jin et al., 1997; Menning, 1995), major sequence-bounding unconformities and formations in proximal basin margin areas and correlative units in distal basin centre areas, assumed (Golonka et al., 1994) and measured (Ogg and Steiner, 1991) palaeolatitudes in °N, on-land climatic setting based on non-marine indicators (Beauchamp, 1994) and palynomorphs (Utting, 2001), and assumed marine temperatures based on biogenic and abiogenic indicators (Beauchamp and Desrochers, 1997). Modified from Beauchamp (1995) and Beauchamp et al. (2001).

tions of northern Alaska (Moore et al., 1994; Beauchamp, 1995), the Hart River and Blackie formations of the northern Yukon Territory (Richards et al., 1997; Beauchamp, 1995), the Banff and the Prophet formations from the southwest regions of the Northwest Territories to southwestern Alberta (Richards et al., 1993), the Lodgepole and Middle Canyon formations and the Wilson Creek Member of the Wood River Formation of the northwestern United States (Pole and Sandberg, 1991; Reid and Dorobek, 1991; Mahoney et al., 1991).

The Bashkirian to Sakmarian Hare Fiord Formation of the Sverdrup Basin is a good example of such deposits (Fig. 1). The Hare Fiord Formation comprises spiculitic limestones associated with dark-brown mudrocks (argillaceous limestones, siltstones, shales and cherts) occurring as thinly- to medium-bedded rythmites of turbiditic origin (Fig. 2; see Thorsteinsson, 1974; Davies, 1977; Beauchamp et al., 1995b). The Hare Fiord Formation is the distal, slope to basinal equivalent to the Nansen Formation and equivalent formations (Fig. 1). These more proximal units form

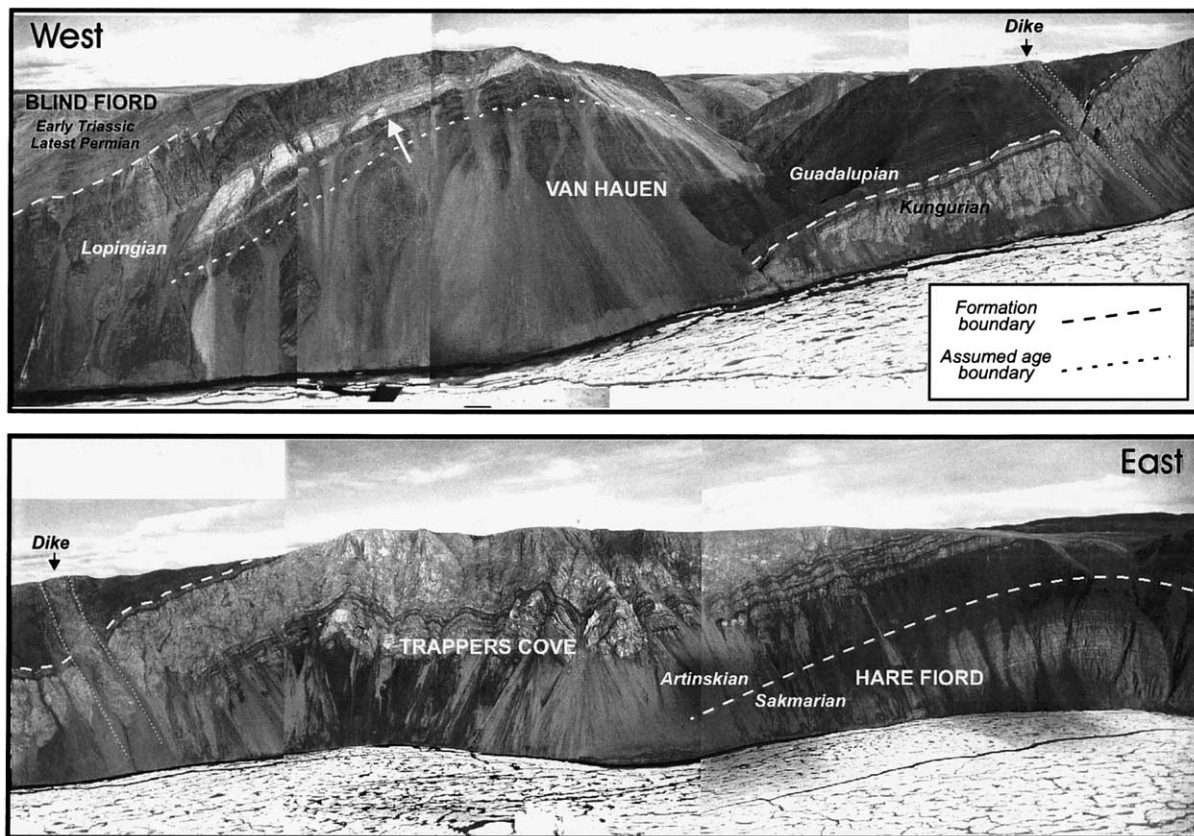


Fig. 2. Permian Chert Event (PCE) west and east of Smith Creek, NW Ellesmere Island. Onset of PCE corresponds to contact between Sakmarian and older Hare Fiord Formation and Artinskian to Kungurian Trappers Cove Formation. Demise of PCE corresponds to contact between Guadalupian and Lopingian van Hauen Formation and latest Permian–Early Triassic Blind Fiord Formation. More than 3 km of coastline is shown. A Cretaceous dike provides a tie between two sets of composite photographs. Hare Fiord Formation comprises 250 m of turbiditic slope mudrocks (variably cherty, argillaceous spiculitic limestones). Trappers Cove Formation is 400 m of basin to slope black siliceous shales passing upward into cliff-forming spiculitic cherts. Van Hauen Formation is composed of 400 m of siliceous shales passing upward into black spiculitic chert. Light-coloured unit (white arrow) high in the van Hauen Formation is a prograding tongue of relatively shallow spiculitic cherts. Blind Fiord Formation comprises chert-free shales and siltstones.

a thick succession of reefal shelf-edge and cyclic shelf carbonates that range from deep- and cool-water wackestones and packstones to shallow- and warm-water packstones and grainstones. The latter accumulated under warm tropical-like conditions as shown by high rates of carbonate production associated with a rich and diversified chlorofoam

biota (Fig. 3; Beauchamp, 1994; Beauchamp and Desrochers, 1997). A wide variety of phylloid algal and *Palaeoaplysina* buildups ranging from small patch reefs to large continuous reef-tracts occur throughout the Bashkirian to Sakmarian succession (Beauchamp et al., 1989a,b,c; Beauchamp, 1993). Marine cements in the form of aragonite botryoids and high-Mg radial crusts are ubiquitous throughout these deposits (Davies and Nasichuk, 1991). Ooids and oncoids are common to abundant.

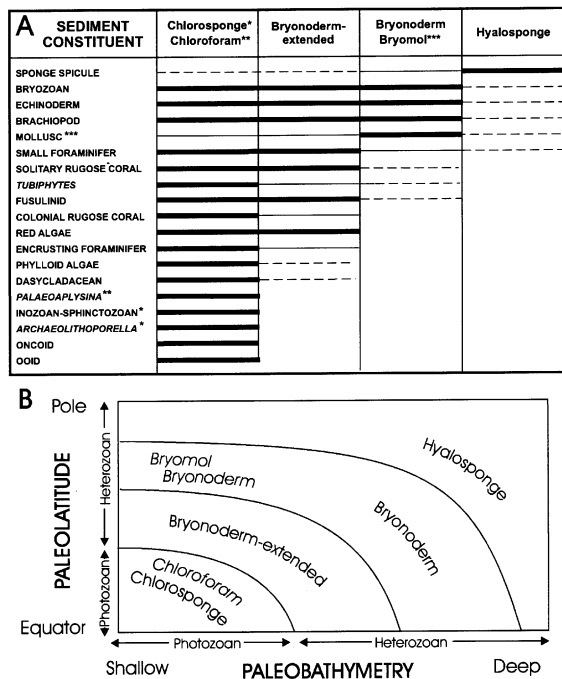


Fig. 3. (A) Biotic composition and associated non-skeletal elements (ooids, oncoids) of warm- (chlorosponge, chlorofoam), cool- and cold- (bryonoderm-extended, bryonoderm, bryomol), and very cold-water (hyalosponge) Late Paleozoic biogenic sediments. Inozoan and sphinctozoan sponges and *Archaeolithoporella* are restricted to chlorosponge assemblage. *Palaeoaplysina* is part of chlorofoam assemblage. Molluscs (bivalves) are abundant only in bryomol assemblage. (B) Hypothetical, mostly temperature-controlled, distribution of warm- to very cold-water assemblages related to water depth and latitude. Modified after Beauchamp (1994) and Beauchamp and Desrochers (1997). Photozoan and heterozoan terminology from James (1997). Photozoan assemblages and associated carbonates formed in warm and shallow environments. Heterozoan assemblages and associated carbonates formed in a variety of shallow- to deep-, cool- to cold-water environments. Other factors such as salinity fluctuations and light deprivation (Lees and Buller, 1972; Lees, 1975) can cause heterozoan sediments to accumulate even in relatively warm water.

Early Permian warm-water chlorofoam carbonates form a several thousand kilometre long belt (Fig. 4A; Table 1), that extends from the pre-Uralian and Timan-Pechora areas of Russia to the southwestern USA. Among other warm-water organisms (Stevens, 1982; Fedorowski et al., 1999), this belt is characterised by the presence of *Palaeoaplysina*, the plate-like hydrozoan(?) that occurs in association with phylloid algae within both reefal and non-reefal successions (Beauchamp et al., 1989a; Breuninger et al., 1989; Ritter and Morris, 1997). *Palaeoaplysina* is not present in the Tethys or in southwestern USA (Fig. 4A) where warm-water carbonates are represented by a more diversified chlorosponge assemblage (Fig. 3; Beauchamp, 1994; Beauchamp and Desrochers, 1997). *Palaeoaplysina*-bearing chlorofoam biota coexists with Tethyan-like chlorosponge biota in the reefal succession of the southern Urals (Vennin, 1997).

2.2. Onset of PCE: Late Sakmarian–Early Artinskian

For Murchey and Jones (1992), massive chert deposition in the deep distal environments of northwest Pangea started at some indeterminate time during the Leonardian, that is during either the Artinskian or the Kungurian. Much of this uncertainty reflects the difficulty of equating Permian biostratigraphic schemes, based on radiolarians, with those based on benthic shelly fossils, or conodonts. In addition, the lack of biostratigraphically-reliable Lower Permian deeper-water spicule-rich sections along the margin of western North America prevented Murchey and Jones (1992) from providing a better estimate of the

onset of the PCE. A far more precise answer comes from the Sverdrup Basin.

Spiculitic chert is the dominant lithology in the Artinskian Trappers Cove Formation which overlies the Sakmarian Hare Fiord Formation in the distal areas (Figs. 1 and 2). Up to 1 km of thinly- and medium-bedded Trappers Cove cherts interfinger with subordinate shales, siltstones and scattered carbonate tongues of turbiditic origin (Beauchamp and Henderson, 1994). These dark-coloured strata, representing slope to basinal sedimentation, pass upward and cratonward into carbonate deposits of the Great Bear Cape Formation (Fig. 5), a non-cyclic succession of cherty, fine-grained carbonates passing upward into

coarser fossiliferous packstones and grainstones. The Great Bear Cape Formation is dominated by sponge spicules, bryozoans, echinoderms and brachiopods (bryonoderm assemblage; see Fig. 3). In addition to this biota, the shallowest-water grainstones in the Great Bear Cape Formation locally contain large fusulinaceans and colonial rugose corals (bryonoderm-extended assemblage of Beauchamp and Desrochers, 1997; see Fig. 3). Carbonate buildups are rare in the Great Bear Cape Formation and form isolated mud mounds comprising bryozoan and soft-bodied sponges as the main constituents (Beauchamp, 1989). Submarine cements are non-existent even in the shallowest-water grainstones. Radial high-Mg cal-

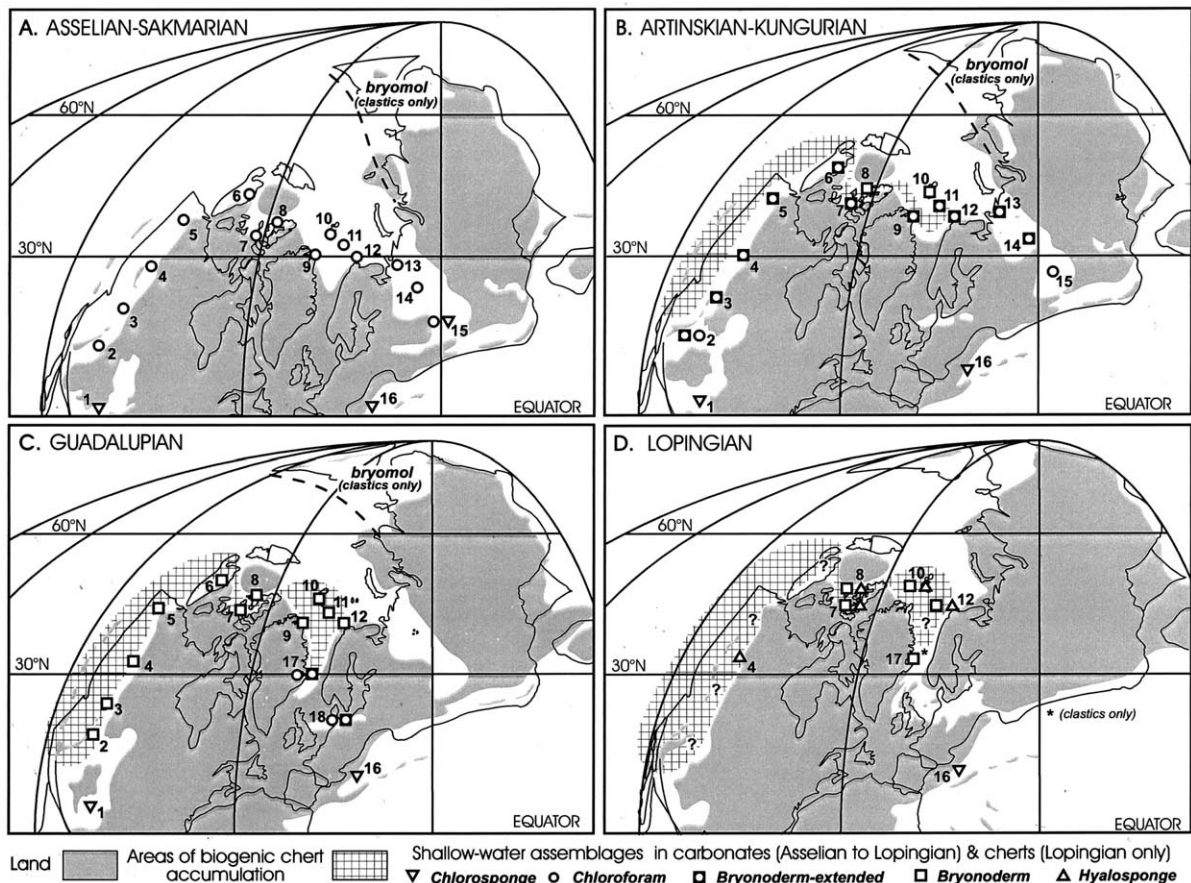


Fig. 4. Permian palaeogeography of northwest Pangea showing land, sea, areas of biogenic chert deposition and shallow-water benthic assemblages. (A) Asselian–Sakmarian. (B) Artinskian–Kungurian. (C) Guadalupian. (D) Lopingian. Paleogeography by Ziegler et al. (1997), modified to display open seaway between Northern Alaska (rotated) and Sverdrup Basin (Embry, 1993), and to show area of marine deposition between North Greenland and Svalbard. Numbered localities are as in Table 1.

cite isopachous layers occur in association with bryozoan-dominated facies in the mud mounds (Beauchamp, 1993, 1989).

The Great Bear Cape Formation represents a distally-steepened, cool-water carbonate ramp that prograded laterally into slope to basinal spiculitic cherts of the Trappers Cove Formation (Beauchamp and Henderson, 1994; Beauchamp and Desrochers, 1997). The entire succession is shallowing upward and forms the Artinskian regressive systems tract (RST) of a Sakmarian–Artinskian T–R sequence (Fig. 5), the transgressive systems tract (TST) of which is represented by the Sakmarian Raanes Formation. The maximum flooding surface (MFS) of the sequence marks the onset of the Permian Chert Event. A conodont assemblage comprising *Streptognathodus inornatus*, *Adetognathus paralautus* and *Mesogondolella bisselli* straddles this surface. This assemblage was originally correlated with the upper part of the Aktastinian substage of the Artinskian (Beauchamp and Henderson, 1994). However, fusulinaceans recovered from within the Raanes Formation at Grinnell Peninsula, Devon Island, have indicated a Sakmarian age for an interval that also contains *S. inornatus* and *A. paralautus* (Beauchamp et al., 1998). Accordingly, Henderson and Mei (2000) and Mei and Henderson (2001) now roughly correlate the Raanes–Great Bear Cape contact with the Sakmarian–Artinskian boundary, which is the age interpretation followed herein and corroborated by other fossil groups such as colonial corals, ammonoids, small foraminifers and fusulinaceans (Beauchamp et al., 1998). The onset of the PCE is hence dated at or very near the Sakmarian–Artinskian boundary (Fig. 5). In addition to the maximum flooding, this event coincides with the ending of high-frequency/high-amplitude shelf cyclicity, the onset of massive biogenic chert deposition in the deep-water distal areas, and a long-term shift from warm- to cool-water carbonate sedimentation in the shallow-water proximal areas (Fig. 1). A coeval shift to cool-water carbonates in the proximal areas and massive chert deposition in the distal areas can be traced from Svalbard and the Barents (Fig. 4B, Table 1) to the

Nevada–Idaho border to the south (Riepetown Formation; Murchey and Jones, 1992).

2.3. Expansion of PCE: Guadalupian

More than 1 km of late Early, Middle and Late Permian cherts belonging to the van Hauen Formation occur in the distal areas of the Sverdrup Basin (Fig. 1). The van Hauen Formation (Thorsteinsson, 1974) is composed of dark-coloured spiculitic cherts (Fig. 2), associated with minor turbiditic limestones, siltstones and shales that accumulated in a distal, slope to basinal setting (Desrochers and Beauchamp, 1995). Age-diagnostic conodonts (Beauchamp et al., 1989b), brachiopods (Thorsteinsson, 1974), palynomorphs (Utting, 1994) and one ammonoid (Nassichuk, 1995) have been recovered in the van Hauen Formation and equivalent units. The van Hauen Formation is a continuous sedimentary unit that records distal deposition of three broad sequences that are bounded by major unconformities in the proximal areas (Fig. 1; Beauchamp et al., 1989b,c; Beauchamp and Thériault, 1994; Beauchamp, 1995). The middle sequence of Roadian–Wordian age shallows up and passes landward into the correlative Degerbøls Formation, which is composed of cherty limestones dominated by sponge spicules, bryozoans, brachiopods and echinoderms (bryonoderm assemblage; see Fig. 3), associated with minor solitary rugose corals (*Calophyllum* fauna of Fedorowski and Bamber, 2001), frondicularid foraminifers and rare molluscs. The formation is devoid of fusulinaceans, colonial corals, calcareous algae or reefs. There is no submarine cement even in the shallowest water facies represented by cross-bedded, high-energy crinoidal grainstones. In contrast, glauconite is an important constituent and is the dominant mineralogical species in the nearshore equivalent Trold Fiord sandstones (Fig. 1), which also locally contain some dropstones (Beauchamp, 1994). The impoverished biotic composition of the Degerbøls carbonates, the lack of marine cement, and the presence of dropstones (floating ice) and of glauconite (slow sedimentation rate) indicate that the Sverdrup Basin area was bathed by

Table 1
 Permian shallow-water biotic assemblages in carbonates and cherts reported from selected areas and stratigraphic units of northern Pangea

| Areas | Asselian-Sakmarian | Artinskian-Kungurian | Guadalupian | Lopingian | References |
|--|---|--|--|---|--|
| 1. Permian Basin | chlorosponge, Hueco Group (Hueco Mts.) | chlorosponge, various units (Glass Mts.) | chlorosponge, Capitan (Guadalupe Mts.) | marine evaporites | Babcock, 1977; Wahlman, 1985, 1988, in press |
| 2. NW United States, Phosphoria Basin | chlorofoam, various units | bryonoderm-extended, Trapper Creek, chlorofoam, Pequoop | bryonoderm, Phosphoria | missing | Wardlaw, 1980; Mytton et al., 1983; Wardlaw and Collinson, 1986; Breuninger et al., 1989; Wardlaw et al., 1995 |
| 3. SW Alberta | chlorofoam, Telford | bryonoderm-extended, Ross Creek, Johnston Canyon | bryonoderm, Ranger Canyon | missing | Henderson et al., 1993; Richards et al., 1993 |
| 4. NE British Columbia, SW N.W.T. | chlorofoam, Belcourt, Belloy | bryonoderm-extended, Belcourt, Belloy | bryonoderm, Fantasque | hyalosponge | Henderson et al., 1993; Richards et al., 1993; Wang et al., 1994 |
| 5. Yukon Territory | chlorofoam, Jungle Creek | bryonoderm-extended, Jungle Creek, Tahkandit | bryonoderm, Sadlerochit, Tahkandit, unnamed unit | missing | Bamber and Waterhouse, 1971; Henderson et al., 1993; Beauchamp, 1995; Richards et al., 1997 |
| 6. Alaska's North Slope | chlorofoam, 'transitional' Lisburne | bryonoderm-extended, 'transitional' Lisburne | bryonoderm, Echooka (Sadlerochit) | missing | Detterman et al., 1975; Dutro and Silberling, 1988; Crowder, 1990; Moore et al., 1994; Beauchamp, 1995 |
| 7. Sverdrup Basin, southern margin | chlorofoam, Nansen, Belcher Channel, Tanquary, Raanes | bryonoderm-extended, bryonoderm, Great Bear Cape | bryonoderm, Degerbols | bryonoderm, hyalosponge, unnamed unit | Thorsteinsson, 1974; Nassichuk and Wilde, 1977; Beauchamp, 1993, 1994, 1995; Beauchamp and Henderson, 1994; Beauchamp and Desrochers, 1997 |
| 8. Sverdrup Basin, northern margin | chlorofoam, Nansen, Raanes | bryonoderm, Great Bear Cape | bryonoderm, Degerbols | bryonoderm, hyalosponge, unnamed unit | Thorsteinsson, 1974; Davies and Nassichuk, 1991; Beauchamp, 1993, 1994, 1995; Beauchamp et al., 1995a,b; Beauchamp and Henderson, 1994; Beauchamp and Desrochers, 1997 |
| 9. North Greenland | chlorofoam, Kim Fjelde | bryonoderm-extended, bryonoderm, Kim Fjelde (upper part) | bryonoderm, Midnatsfjeld | missing | Stemmerik and Håkansson, 1989; Stemmerik and Worsley, 1995; Stemmerik, 2000 |
| 10. Svalbard | chlorofoam, Gypsdaalen Group | bryonoderm, Vøringen (Kapp Starostin) | bryonoderm, Kapp Starostin | bryonoderm, hyalosponge, Kapp Starostin | Gruszczynski et al., 1989; Stemmerik and Worsley, 1989; Ezaki et al., 1994; Stemmerik et al., 1995; Stemmerik, 1997, 2000 |
| 11. Bjornøya | chlorofoam, Kapp Duner | bryonoderm-extended, bryonoderm, Hambergfjellet | bryonoderm, Miseryfjellet | missing | Stemmerik and Worsley, 1989, 1995; |
| 12. Finnmark Platform, Barents Shelf, (subsurface) | chlorofoam, unnamed units | bryonoderm-extended, bryonoderm, unnamed unit | bryonoderm, unnamed unit | bryonoderm, hyalosponge, unnamed unit | Bugege et al., 1995; Stemmerik and Worsley, 1995; Blendinger et al., 1997; Ehrenberg et al., 2000 |

Table 1 (Continued).

| Areas | Asselian-Sakmarian | Artinskian-Kungurian | Guadalupian | Lopingian | References |
|---------------------|---|--|---|--|--|
| 13. Timan-Pechora | chloroform, various units | bryonoderm-extended, various units | marine clastics | marine and continental clastics | confidential IKU reports, Zhemchugova and Schamel, 1994; Wahlman and Konovalova, in press; Wahlman, in press |
| 14. Central Urals | chloroform, Shikhansky and Tastubsky horizons (Belaya Gora) | bryonoderm-extended, Burtsevsky Horizon (Belaya Gora), Kamaiskaya and Sylvinskaya suites (Kungur area) | marine evaporites, clastics | continental clastics | Ekhlakov, 1991; Ozhigbesov et al., 1991; Beauchamp, in preparation |
| 15. Southern Urals | chloroform, Tratau-Shaktau shikhans | chloroform, Tratau-Shaktau shikhans | marine evaporites, clastics | continental clastics | Chuvashov, 1983; Vennin, 1997 |
| 16. Western Tethys | Trogkofel (Carnic Alps) | chlorosponge, Trogkofel (Carnic Alps) | chlorosponge, Bled Reef (Slovenia) | chlorosponge, Skiathos Series (Skyros) | Flügel, 1981; Flügel et al., 1984; Flügel and Reinhardt, 1989; Wahlman, in press |
| 17. East Greenland | continental clastics | missing | bryonoderm-extended, Karstryggen, Wegener Halvø | bryonoderm (*) Schuchert Dal (*) from a clastic unit | Surlyk et al., 1986; Scholle et al., 1991; Stemmerik and Worsley, 1995 |
| 18. Zechstein Basin | continental clastics | continental clastics | bryonoderm-extended, Zechstein 1-3, magnesian limestone | evaporites | Kiersnowski et al., 1995; Smith, 1981a,b |

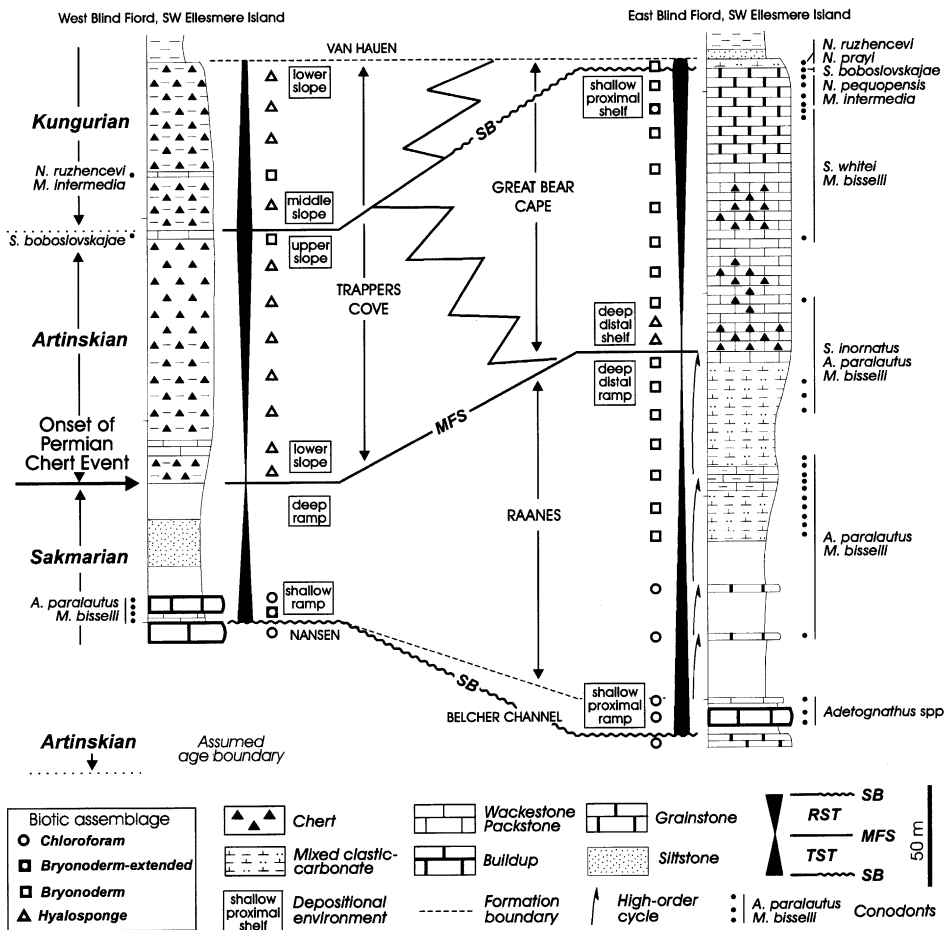


Fig. 5. Measured sections west and east of Blind Fiord, SW Ellesmere Island, Canadian Arctic, recording onset of Permian Chert Event. Diagram shows lithologies, formations, conodont assemblages, sequence stratigraphic interpretations, depositional environments, biotic assemblages and age interpretations. West Blind Fiord section is type section of Trappers Cove Formation. East Blind Fiord section is type section of Raanes and Great Bear Cape formations. Modified from Beauchamp and Henderson (1994).

cold water during the Guadalupian (Beauchamp, 1994; Beauchamp and Desrochers, 1997).

Evidence for a Guadalupian landward expansion of silica factories and a coeval contraction of carbonate factories can be found all along the western margin of North America (Fig. 4C). At one end of the spectrum, siliceous deposits comprise thinly-bedded distal, deep-water basinal to slope radiolarian chert found in several exotic terranes from Alaska to California (Murchey and Jones, 1992). Some of these units contain sediments deposited next to Panthalassan seamounts and island arcs that lay offshore of North Amer-

ica during the Middle and Late Permian and were later accreted to the craton. Other units comprise thicker-bedded, lighter-coloured, bioturbated shallower water shelf to ramp spiculitic chert deposited along the western margin of North America. Among those are the Rex Chert and Tosi Chert members of the Phosphoria Formation in the western USA (Wardlaw and Collinson, 1986), which interfinger with fossiliferous, cold-water, bryonoderm carbonates with echinoderms, bryozoans, brachiopods and sponge spicules as the only constituents (Wardlaw and Collinson, 1986). Similar and coeval units include the

chert-bearing Fantasque Formation of western Canada (Henderson et al., 1993), and Kapp Starostin Formation of Svalbard (Stemmerik and Worsley, 1989) and equivalent units in the Barents Sea (Stemmerik and Worsley, 1995).

There is no Guadalupian warm-water chloroform-type carbonate along the western margin of North America. Yet chlorosponge warm-water carbonates are ubiquitous in West Texas (Fig. 4C). Guadalupian warm-water carbonates also accumulated in East Greenland where they are represented by a variety of karstified stromatolitic and oolitic facies that belong to the Karstryggen and Wegener Halvø formations (Surlyk et al., 1986). Carbonate buildups that formed in a more offshore setting in the Wegener Halvø Formation comprise a typical bryonoderm biota which probably indicates a deeper-water, not necessarily colder, environment. However, the Ravnfeld Formation, a deeper-water mudrock unit laterally equivalent to the Wegener Halvø buildups, contains abundant spiculitic cherts (Surlyk et al., 1986).

2.4. Zenith of PCE: Lopingian

The Lopingian record of northwest Pangea is poorly understood due to the lack of diagnostic fossils and patchy preservation beneath the sub-Triassic unconformity. Yet, the Lopingian, at least its early part, is believed to coincide with the zenith of the PCE, as shown by chert-rich strata worldwide. In the Sverdrup Basin, a widespread transgression occurred during the Capitanian leading to the deposition of bryonoderm carbonates and spiculitic cherts in the upper part of the Degerbøls and Trolld Fiord formations and dark-coloured spiculite in the correlative van Hauen Formation (Fig. 1). A major regression followed, the onset of which is roughly correlated with the Guadalupian–Lopingian boundary (Fig. 6). Distal deep-water cherts are represented by dark-coloured spiculites in the upper part of the unit currently mapped as van Hauen Formation (Fig. 2). These sediments pass upward and landward into pale grey to white spiculitic chert, the more proximal and shallower accumulations of which are either extensively bioturbated or

cross-bedded and contain variable amounts of glauconite (Beauchamp et al., 1995a; Beauchamp and Desrochers, 1997). Spiculitic chert (hyalosponge assemblage; see Fig. 3) is the dominant lithology in the distal to proximal Lopingian RST. Rare bryonoderm carbonates occur as pods and lenses. This shallower-water chert is part of a yet unnamed formation containing long-ranging conodonts (Mei and Henderson, 2001). Sedimentary environments represented by the assumed Lopingian chert in the Sverdrup Basin range from distal, deep-water basin and slope to relatively shallow, gently dipping ramp. Sedimentation rates were low as shown by the abundance of glauconite near the basin margin.

As in the Sverdrup Basin, the Late Permian of the Svalbard–Barents Sea area has traditionally been interpreted as missing beneath a widespread sub-Triassic unconformity (see Stemmerik and Worsley, 1989). However, physical and geochemical evidence of Lopingian sedimentation and continuous deposition across the Permian–Triassic boundary has been documented in Svalbard (Gruszczynski et al., 1989; Wignall et al., 1998). The Kapp Starostin Formation is now seen as passing conformably into the overlying Triassic Vardebukta Formation in the more distal areas (Bugge et al., 1995). Likewise, an unnamed pale-coloured chert unit encountered in the subsurface of the Finnmark platform (Bugge et al., 1995; Ehrenberg et al., 2000) is identical, stratigraphically and compositionally, to the yet unnamed pale-coloured Late Permian chert unit in the Sverdrup Basin (Fig. 4D). The same applies for western Canada where the upper part of the Fantasque chert could range into the Lopingian (Table 1, Henderson and Mei, 2000) and may be overlain conformably by Early Triassic mudrocks of the Spray River Group (Wang et al., 1994). Lopingian strata are missing beneath the Early Triassic Dinwoody Formation in the Phosphoria Basin of the western USA (Wardlaw et al., 1995), but Lopingian cherts most occur widely in the upper part of the various allochthonous and exotic units from Alaska to California (Cordey, 1998; Cordey et al., 1999).

The only well-dated Lopingian unit along the northwest margin of Pangea is the Schuchert Dal

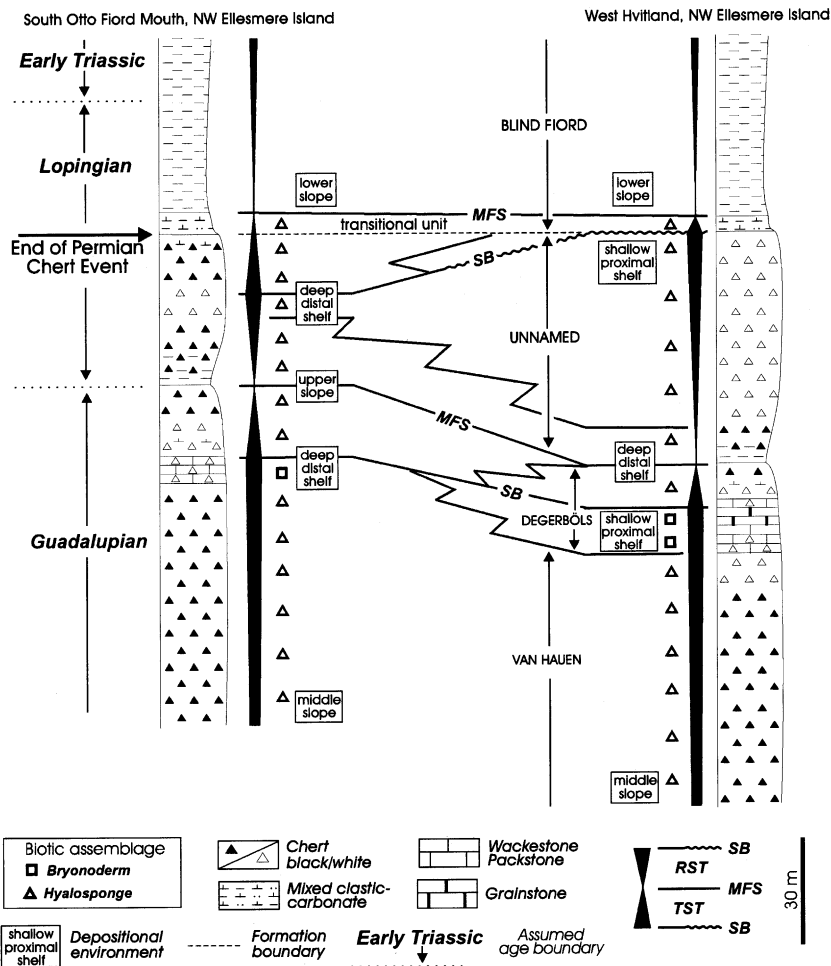


Fig. 6. Measured sections south and north (West Hvitland) of mouth of Otto Fiord, NW Ellesmere Island, Canadian Arctic, recording end of Permian Chert Event. Diagram shows lithologies, formations, sequence stratigraphic interpretations, depositional environments, biotic assemblages and age interpretations. Early Triassic–Lopingian age boundary placed 32 m above base of Blind Fiord Formation at level assumed to roughly coincide with first appearance of conodont *Hindeodus parvus*, based upon data from Otto Fiord South section, 60 km east of West Hvitland section, as documented by Henderson and Baud (1997). Lopingian–Guadalupian boundary placed arbitrarily at contact between Degerbøls Formation and overlying unnamed chert unit.

Formation of East Greenland which has yielded diagnostic ammonoids (Nassichuk, 1995). The Schuchert Dal Formation comprises marine sandstones (Bredehorn Member), which is locally calcareous (Depot Ø Member), and passes laterally into deeper-water mudrocks (Oksedal Member). Biota within the Schuchert Dal Formation forms a low diversity bryonoderm-like assemblage of bryozoans, brachiopods, ammonoids (*Cyclolobus*, *Paramexioceras*), small foraminifers and solitary

rugose corals (Birkelund and Perch-Nielsen, 1976; Surlyk et al., 1986; Pattison and Stemmerik, 1996).

Guadalupian and Lopingian radiolarian cherts also occur in Japan (Murchey and Jones, 1992; Ishiga and Yamakita, 1993; Isozaki, 1994, 1997), representing deep, offshore pelagic Panthalassan sedimentation. Late Permian chert is also common in the Tethyan regions (Marcoux and Baud, 1986; Baud et al., 1991).

2.5. Demise of PCE: latest Permian

Biogenic silica factories collapsed during Late Permian time throughout the world. In the Sverdrup Basin, the end of the PCE coincides with the contact between the chert-rich van Hauen Formation (or equivalent proximal chert unit) and the chert-free Blind Fiord Formation (Figs. 1 and 2). Contrary to a long-established belief (Nassichuk et al., 1973; Thorsteinsson, 1974), there is no physical unconformity between these two units in the central and distal parts of the basin, where non-siliceous greenish grey shales of the Blind Fiord Formation lie conformably upon black siliceous shales of the van Hauen Formation. Evidence of erosional truncation does occur in the more proximal areas as the Blind Fiord Formation lies upon progressively older Permian units towards the basin margin. The base of the Blind Fiord Formation in marginal areas recorded the culmination of a major transgression that led to the drowning of previously submerged, shallow- to deep-water basin axial areas and to the flooding of previously emerged areas well beyond the former basin margin. An equally significant and correlative transgression occurred throughout western and northwestern Pangea and units compositionally similar to the Blind Fiord Formation occur from Idaho to the Barents Sea (Baars, 1988; Mørk et al., 1989; Gibson, 1993).

For many years the base of the Blind Fiord Formation was viewed as marking the base of the Triassic period. Four stages (Griesbachian, Dienerian, Smithian, Spathian) defined in the Sverdrup Basin were, until recently, recognised as the standards for Early Triassic time (Tozer, 1967). However, the base of the Blind Fiord Formation, and hence its conformable contact with the van Hauen Formation in the distal parts of the Sverdrup Basin, is now interpreted as lying within the latest Permian Changhsingian stage. This is based on the occurrence of *Hindeodus parvus* recovered 32 m above the base of the Blind Fiord Formation in a section south of the head of Otto Fiord, NW Ellesmere Island (Henderson and Baud, 1997). The conodont *H. parvus* has been accepted by the International Commission on

Stratigraphy as the marker for the Permian–Triassic boundary with a stratotype at Meishan, China (Yin et al., 2001). At Otto Fiord South, this key conodont taxa lies above a typical, uppermost Permian association with *Neogondolella* sp. cf. *subcarinata* and *Neogondolella* sp. aff. *changhsinensis* from within the lower part of the Blind Fiord Formation, which also contains the ammonoid *Otoceras concavum* (Henderson and Baud, 1997). As a result, the demise of the silica factories, not only in the Sverdrup Basin, but around the world is referred to as an ‘end-Permian’ event or a ‘latest Permian’ event, as opposed to a ‘P–T boundary’ event (Fig. 6). The same applies to a number of other biological, geochemical and environmental events that accompanied the transition from the Palaeozoic to the Mesozoic (Baud, 2001). These events invariably occurred prior to the first appearance of *H. parvus* (Bowring et al., 1998; Kozur, 1998a,b).

In the most distal parts of the Sverdrup Basin, the end of the PCE is recorded in the sharp, yet conformable, contact between black shale of the van Hauen Formation and greenish grey shale of the Blind Fiord Formation (Fig. 1). The upper 100 m of the van Hauen Formation is deepening upward and transgressive. It comprises medium-bedded black spiculitic chert passing upward into increasingly argillaceous and thinly-bedded siliceous shale. The Blind Fiord Formation is broadly shallowing upward and regressive. It comprises very thin, finely-laminated shale, passing upward into increasingly shallow turbiditic siltstone and fine-grained sandstone. The contact between the two formations hence roughly coincides with an MFS.

The end of the PCE is recorded in a possibly less condensed transitional unit between the Blind Fiord Formation and underlying, unnamed spiculitic chert unit south and north of the head of Otto Fiord, NW Ellesmere Island, which lies in a more proximal area of the Sverdrup Basin (Fig. 6). The transitional unit comprises 4–6 m of finely-laminated, pyrite-rich siltstone interfingering with carbonate. The latter, composed of ankerite, occurs as lenses or continuous beds 1–2 cm in thickness. In thin sections, the carbonates are micritic and display the ghost of scattered to

abundant dissolved sponge spicules. Pyrite is abundant and occurs as framboids lying along fine laminations in the sediment. Microcrystalline silica is a very minor component as it locally fills the void left by dissolved spicules. The exquisite preservation of the laminations indicates that the sediment was undisturbed by bioturbation. It is believed to represent a rare record of the final phase of the transgression that preceded deposition of the Blind Fiord shale. More importantly it recorded the transition between a chert-rich to a chert-free Sverdrup Basin.

2.6. *Early Triassic Chert Gap: Griesbachian to Spathian*

Spiculitic chert deposition did not resume in the various sedimentary basins and continental slopes along the northwest margin of Pangea where the Mesozoic succession is dominated by clastic sediments. In the Sverdrup Basin, Early Triassic clastic sediments are contained in the Blind Fiord Formation (Figs. 1 and 6), which comprises shales, siltstones and fine-grained sandstones, representing distal sedimentation within an array of environments that ranged from basin, slope to deep clastic shelf (Embry, 1991). Blind Fiord shales comprise only scant fossils (molluscs, bryozoans, ammonoids), the assemblage of which represents post-extinction biota. Red calcareous algae have been reported from the Vardebukta Formation of western Spitsbergen (Wignall et al., 1998; Nakrem and Mørk, 1991), which is correlative and compositionally similar to the Blind Fiord Formation (Mørk et al., 1989). In the Sverdrup Basin, the entire Mesozoic was a time of high clastic influx and carbonates are exceedingly rare throughout the succession, while bedded cherts, cherty mudrocks or even chert nodules are absent.

The Blind Fiord Formation passes cratonward into the Bjerne Formation (Fig. 1), which comprises sandstones associated with minor shales and conglomerates that were deposited in a variety of environments ranging from shallow shelf, nearshore subtidal, shoreface to non-marine (Embry, 1991). Fluvial overbanks in the Bjerne Formation are invariably red-coloured and associated

with caliche palaeosols, which does indicate a return to a more arid, and presumably warmer, setting during the Early Triassic, a suggestion corroborated by the palynological record (Utting, 1994, 2001).

There is no Griesbachian and Dienerian chert in the rock record. In Japan, where the oldest record of Triassic chert is preserved in a deep oceanic setting (Isozaki, 1994), deposition of siliceous claystone resumed during the Smithian, but *bona fide* radiolarian and spiculitic chert reoccurred only during the Late Spathian (late Early Triassic) or the Anisian (early Middle Triassic), that is after a 8–10 Ma interruption (Isozaki, 1997; Kakuwa, 1996; Suzuki et al., 1998). The gap appears the same in the Cache Creek Terrane of British Columbia (Isozaki, 1997; Cordey, 1998). Radiolarian chert deposition resumed only during the Late Triassic in the deep oceanic record of Alaska (Murchey et al., 1988).

3. Discussion and implications

3.1. *Formation and preservation of biogenic chert*

Six fundamental conditions have to be met for biogenic siliceous sediments to precipitate, accumulate and be preserved as bedded chert in the rock record. These are: (1) an adequate supply of silica and nutrients into the world oceans (Laschet, 1984; Racki, 1999), (2) a mechanism for bringing silica and nutrients to the site of biogenic productivity and maintaining a steady supply thereof (Wilde et al., 1990; McGowran, 1989), (3) an opportunistic pelagic and/or benthic silica-secreting biota capable of taking advantage of an adequate supply of silica and nutrients (Hein and Parrish, 1987; Racki, 1999), (4) a suitable marine environment for this biota to flourish (Hein and Parrish, 1987; Racki, 1999), (5) a favourable post-mortem setting in the water column, on the sea-floor and in the shallow burial environment to minimise silica dissolution (Laschet, 1984; Nelson et al., 1995), and (6) minimal dilution with carbonate and/or clastic sediments to ensure purity of the chert facies (Laschet, 1984). Conditions 1–4 dictate the extent of silica

precipitation. Condition 5 controls the degree to which biogenic silica will withstand dissolution and accumulate as sediments. Condition 6 determines whether pure chert or a mixture of various lithologies will enter the burial environment. Major fluctuations in any one of these conditions over geological times could have led to either chert-rich or chert-poor intervals (Hein and Parrish, 1987).

Modern oceans are undersaturated with respect to silica due to biological activity. Because opaline silica is unstable (Laschet, 1984), at least 50% of the silica produced by diatoms in the euphotic zone dissolve in the upper 100 m of the water column (Nelson et al., 1995). Chemical and physical weathering of the continents provides the main source of dissolved and particulate silica, which is transported as river runoff (83.5%) and aeolian dust (7.5%) (Treguer et al., 1995). Only a minor portion of marine silica (3.0%) comes from the Earth's interior through submarine venting and other hydrothermal processes associated with volcanic centres, whereas twice as much (6.0%) silica is derived from the submarine weathering of sea-floor basalts (Treguer et al., 1995; Harrison, 2000). According to Racki (1999), secular episodes of unusually high chert deposition are likely to represent times of increased supply of silica in the world oceanic system, due to increased weathering and/or volcanism. Similarly, Gammon et al. (2000) invoked a local riverine influx of nutrients to explain the presence of Eocene spiculitic cherts in southern Australia. A simple application of these ideas would be that the PCE was a time of increased input of silica and/or nutrients into the world oceans, whereas the ETCG coincides with a time of reduced input. This hypothesis finds little support in the geological record.

There is no reason to suggest that increased volcanism triggered the onset of the PCE during the Late Sakmarian–Early Artinskian interval, nor that the Early Triassic was a time of reduced volcanism. The $^{87}\text{Sr}/^{86}\text{Sr}$ curve of marine sediments does show a significant late Early to Middle Permian trend towards radiogenic values (Denison and Koepnick, 1995; Martin and Macdougall, 1995), which may suggest an increase in volcanic input, but may equally indicate a de-

crease in continental contribution. However, this trend sharply reverses near the Middle/Late Permian boundary at a time when chert deposition appears to have increased worldwide. Likewise, the end-Permian massive eruption of the Siberian traps (Campbell et al., 1992; Renne et al., 1995) coincides with the collapse, not the outburst, of chert factories around the world.

There is also no reason to believe that continental weathering increased dramatically across the Sakmarian–Artinskian boundary and triggered the onset of the PCE. Episodic waxing and waning of Gondwana glaciers until the Sakmarian (Dickins, 1996) must have led to significant continental weathering and a steady flux of silica and nutrients to the oceans before the PCE (Froelich et al., 1992; Harrison, 2000). Likewise, the Late Carboniferous and Permian weathering of the Hercynian–Alleghanian and Ural mountains must have provided a steady supply of dissolved silica and nutrients to the oceans (Ruddiman and Prell, 1997), both before and after the onset of the PCE. The presumed increase in atmospheric CO_2 often invoked as a cause for the end of Gondwana glaciations (Berner, 1990, 1991; Frakes et al., 1992) would have led to increased continental weathering. In contrast, the demise of Gondwana glaciers coupled with the final consolidation of Pangea would have led to reduced continental weathering as increasingly arid areas were created (Herring, 1995; Barron and Fawcett, 1995; Francis, 1994; Crowley, 1994) and a mega-monsoonal climate was established (Parrish, 1993). Thus, it is not clear to what extent the PCE corresponded with an increase in weathering and flux of continental silica and nutrients to the oceans. As discussed before, silica-secreting organisms (radiolarians, sponges) flourished both before and after the onset of the PCE. What the PCE truly brought was a major increase in the preservation of bedded chert, which may or may not have been associated with a significant increase in productivity.

3.2. Permian cooling

The onset of the PCE in the deeper-water distal areas coincides with a shift from highly diversi-

fied, warm-water carbonates to faunally-impo- verished cool- and cold-water carbonates in the shal- low proximal areas. This shift is a clear indication that much colder marine conditions were estab- lished at about the Sakmarian–Artinskian bound- ary along the northwest margin of Pangea. To some extent, this cooling can be associated with the northward migration of Pangea during the Late Palaeozoic. Various reconstructions (Golon- ka et al., 1994; Scotese and Langford, 1995; Zieg- ler et al., 1998; Golonka and Ford, 2000) show that the Sverdrup Basin migrated from about 25°N during the Late Carboniferous to about 40°N during the Early Triassic (Fig. 1). Palaeo- magnetic measurements performed on the Blind Fiord Formation have indicated a palaeolatitude of about 38°N during the Early Triassic (Ogg and Steiner, 1991). Non-marine climatic indicators support the inference that the Sverdrup Basin traveled from a semi-arid subtropical zone to a more humid, subtemperate zone during the Kun- gurian (Ziegler et al., 1997). Accordingly, lower Kungurian and older fluvial overbank deposits are invariably red and associated with nodular to massive caliches, whereas their upper Kungur- ian and Roadian depositional counterparts are black, coaly shales without pedogenic nodules (Thériault and Desrochers, 1994; Beauchamp, 1994; Beauchamp et al., 2001). Likewise, evapor- ite deposits and pseudomorphs after halite and gypsum occur throughout the Artinskian and old- er succession, but are absent in the upper Kun- gurian and younger Permian succession. The com- position of palynoflora also suggests a shift to a more humid climatic setting across the Artin- skian–Kungurian boundary (Utting, 1994; Beau- champ et al., 2001), and a return to dryer condi- tions during the Wordian and possibly Capitanian (Utting, 2001).

The disappearance of warm-water carbonates at about the Sakmarian–Artinskian boundary seems to predate, by as much as 5–10 million yr, the passage from subtropical to subtemperate latitudes (Fig. 1). More puzzling is the establish- ment of apparently much colder marine condi- tions during the Guadalupian and yet again dur- ing the Lopingian, at a time when the Sverdrup Basin probably lay at about 35°N. These condi-

tions prevailed all along the northwest margin of Pangea, as far south as northern Nevada, which lay at a palaeolatitude not exceeding 15°N. Clearly there appears to be a discrepancy between the non-marine and the marine Permian record of northwest Pangea. While the former supports the palaeolatitudinal interpretation, at least up until the Kungurian, the latter suggests that the area was under the influence of cold to very cold waters. These findings therefore suggest that it is an influx of cold, northerly-derived waters along the margins of northwest Pangea, independent of on-land climates, that led to the onset, expansion and zenith of the PCE.

3.3. Thermohaline circulation

Oceanic upwelling has been invoked to explain the presence of shallow- and cool-water carbonate facies along the western margin of North Amer- ica. For instance, upwelling accompanied by a landward surge of the thermocline was the expla- nation provided by Brandley and Krause (1994, 1997) to account for the formation of ikaite con- cretions and relatively shallow- and cool-water carbonates offshore of warm-water carbonates in the Viséan succession of western Canada. Like- wise, cold oceanic upwelling was the explanation offered by Murchey and Jones (1992) for the ori- gin of the PCE and its association with phos- phate-rich deposits along western North America. Accordingly, eastward-directed upwelling, result- ing from Coriolis-driven wind-forcing along the western continental margin of Pangea, is a plau- sible explanation, since the area was in the north- ern hemisphere from the Early Carboniferous on- ward (Scotese and Langford, 1995). A northern gyre would have also generated a southbound sur- ficial current akin to the modern California Cur- rent and would have brought cold waters all along northwest Pangea (Fig. 7A). However, since background upwelling and surficial circulation re- mained more or less the same throughout the Carboniferous, Permian and Triassic periods (Go- lonka et al., 1994), the introduction of much cold- er waters from the Artinskian onward calls for an additional northern source.

It is the thermohaline circulation of northerly-

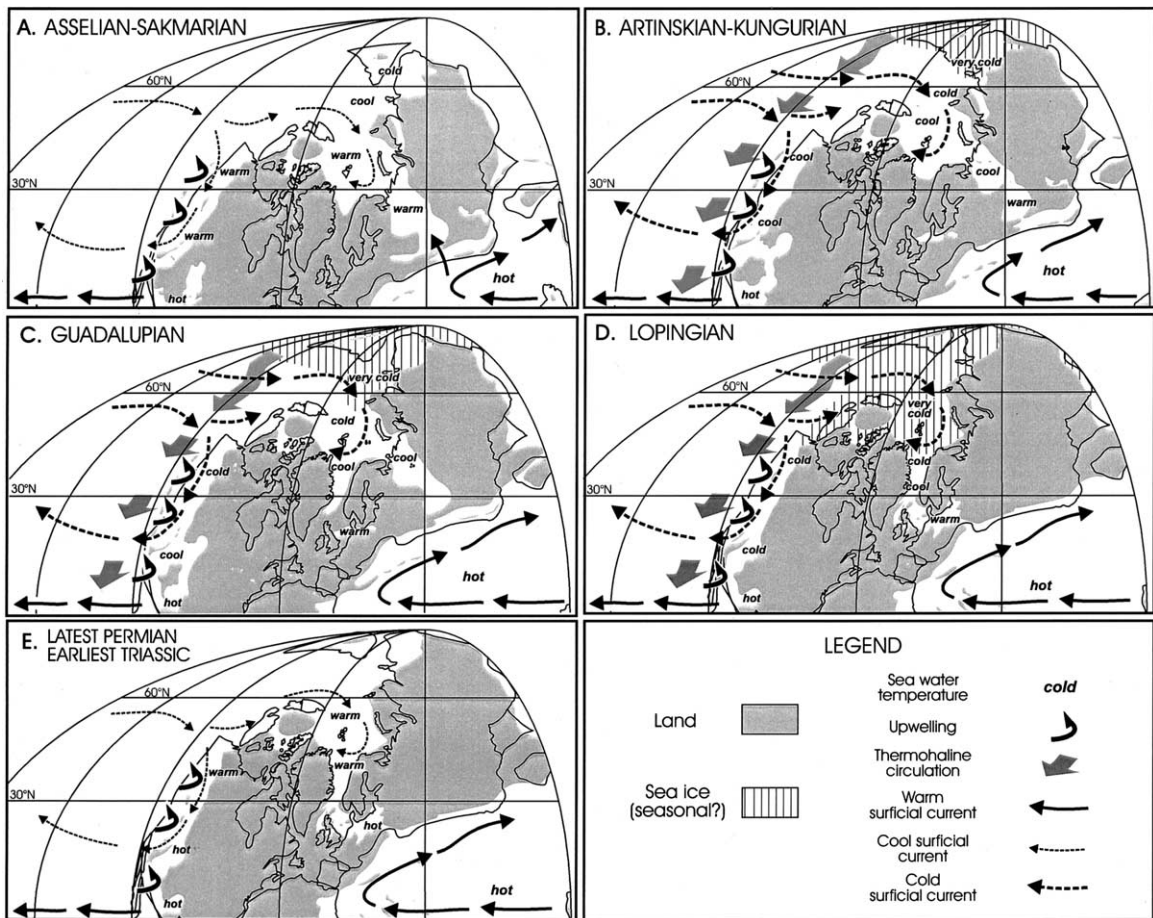


Fig. 7. Palaeogeographic maps of northwest Pangea showing land, sea, extent of assumed sea ice, patterns of surficial warm, cool and cold circulation, upwelling, thermohaline circulation, and interpreted seawater temperatures. Paleogeography modified after Ziegler et al. (1997) to display open seaway between Northern Alaska (rotated) and Sverdrup Basin (Embry, 1993), and to show area of marine deposition between northern Greenland and Svalbard. (A) Asselian–Sakmarian. (B) Artinskian–Kungurian. (C) Guadalupian. (D) Lopingian. (E) Latest Permian–earliest Triassic.

derived nutrient-rich, cold marine waters along the northwestern and western margin of Pangea, in concert with the regional low clastic influx, that likely provided the conditions necessary for the onset, expansion and zenith of the PCE. Given that a sufficient flux of dissolved silica and nutrients from continental weathering and other sources entered the Carboniferous and Permian oceans somewhere (*adequate supply*), and that healthy populations of silica-secreting radiolarians and sponges already existed (*adequate biota*), the introduction of cold waters and its circulation and upwelling along the western and

northwestern margin of Pangea provided a mechanism that replenished the supply of silica in the area (*efficient transportation*), created a nutrient- and oxygen-rich environment favouring siliceous biogenic productivity (*suitable environment*), established cold sea-floor conditions hindering silica dissolution while increasing calcium carbonate solubility (*favourable post-mortem setting*), and provided conditions stressful to biogenic and abiogenic carbonate production (*minimal sediment dilution*), which emptied benthic niches and made them available for colonisation by siliceous sponges.

3.4. Northern sea ice

Invoking thermohaline circulation as a major contributing factor to the PCE leads to a more fundamental question: what was the engine of that circulation and what caused this engine to be set in motion near the Sakmarian–Artinskian boundary, to function increasingly well for the next 30 Ma and to stop abruptly near the end of the Permian?

In the modern oceans, thermohaline circulation in the deep sea is generated by the formation of cold, dense water in the polar regions that then sinks and flows toward the equator, displacing a number of water masses and generating upward movements of nutrient-rich waters (Pinet, 1992). For instance, the North Atlantic deep-water current (NADW), which originates between Greenland and Svalbard as cold melt water from the seasonal thawing of the Arctic ice sheet, sinks and travels from the North to the South Atlantic. Cold and less saline waters originating from the seasonal thawing of Arctic sea ice also flows southward in shallower depths along the eastern Greenland Shelf (East Greenland Current), northward along western Greenland (West Greenland Current) and southward again along the Labrador Shelf (Labrador Current). This influx of Arctic waters causes sea water temperatures to remain low throughout the year and seasonal sea ice to form as far south as Nova Scotia in areas beyond the reach of the Gulf Stream. Blankets of siliceous sponge spicules, in many ways similar to the Lopingian spiculitic deposits of northwest Pangea, currently accumulate in relatively shallow water on the Vesterisbanken Seamount, in the central Greenland Sea (Henrich et al., 1992), an area under sea ice cover half of the year. Spiculitic blankets are also known to form beneath, or near the edge of, permanent sea ice both in the Arctic and Antarctic, where the absence of light, in addition to very cold water temperatures stress most carbonate producers (Bullivant and Dearborn, 1967; Domack, 1988; Van Wagoner et al., 1989).

We suggest that the northwest margin of Pangea was, for nearly 30 Ma during the Permian, bathed by cold waters derived from the seasonal

melting of northern sea ice (Fig. 7B). As with the NADW, melt water would have sunk into the deep north Panthalassa Ocean displacing nutrients- (including phosphorus, nitrogen and iron) and silica-rich waters in its path (Wilde et al., 1990), much of it would have been forced along the margin of northwest Pangea via upwelling. Likewise, some of the lighter, perhaps hyposaline, seasonal melt must have been captured by the clockwise wind-driven northern gyre and would have flowed all along northwest Pangea. This process started near the Sakmarian–Artinskian boundary, intensified throughout the Middle and Late Permian time and ceased suddenly in latest Permian time. It led to oceanic conditions much colder than normally expected from the palaeolatitude of the area (Figs. 7B–D), and hence to a substantial increase in the pole-to-equator temperature gradient. Accordingly, the influence of cold northerly-derived water was felt as far south southern Nevada. If the presence of shallow-water spiculitic blankets and the scarcity of carbonate elements are any indication of an at least seasonal ice cover as in modern environments, then sea ice, at least seasonal in nature, must have existed in the Barents Sea and in the Sverdrup Basin during the Lopingian (Fig. 7D). Accordingly, Ustritsky (1973) and Epshteyn (1981) documented several horizons of dropstones in the Middle Permian of the Kolyma Block in Siberia, suggesting that ice existed at high latitudes. Occurrences of Middle Permian dropstones were also documented in the Sverdrup Basin (Beauchamp, 1994). Noteworthy, cooling and polar ice buildup were also invoked by McGowran (1989) to explain the ‘Eocene Silica Burb’.

The idea that northern sea ice built up from the Artinskian onward challenges some of the most established ideas about global Permian climatic evolution. The widespread occurrence of Late Carboniferous and Early Permian glacial deposits in Gondwana and their replacement by coal measures during the Middle and Late Permian has traditionally been seen as indicative of significant global warming (Visser, 1995; Dickins, 1993, 1996). For many, this observation vindicates a popular CO₂ balance model (Berner, 1990, 1991), whereby Late Permian atmospheric CO₂

levels would have been several times their Early Permian and Late Carboniferous levels, hence resulting in a much warmer Middle and Late Permian world. Based strictly on the distribution of land plants, Rees et al. (1999) modeled an ice-free Wordian Pangea, with warm temperate conditions occurring at high northern latitudes from Taymir Peninsula to northern Alaska (see also Ziegler et al., 1998). However, the idea of high atmospheric CO₂ levels during the Late Permian has been challenged by Ekart et al. (1999) who suggested instead that Permian CO₂ levels were generally low, with a rather sharp decrease during the Late Permian. It is difficult to envision the growth of northern sea ice and the Middle to Late Permian contraction of warm-water carbonate factories around the palaeoequator as indicative of a world that was getting warmer. Likewise, the recent documentation of permafrost features in some of the Late Permian coal measures of Australia (Retallack, 1999) surely indicates that much of Gondwana was still under a cold climatic setting despite the lack of glacial deposits. The occurrence of Cordaites plants (Rees et al., 1999) on lands in proximity to marine sea ice in the northern hemisphere could be indicative of extreme seasonality expected for the Pangean configuration (Parrish, 1993), a unique setting with no modern counterpart.

The buildup of northern sea ice, however, may not be indicative of global cooling (Stanley, 1988) but rather of a set of palaeoceanographic conditions that favoured sea ice formation in the northern hemisphere. The progressive impingement of the northernmost tip of Pangea upon the polar regions from the Artinskian onward could have favoured the creation and deflection of cold-water masses along continental margins (Stanley, 1984, 1988), and promoted the growth of sea ice. The closure of the Uralian trough by Kungurian time must have led to further cooling of the area as warm southerly-derived water was suddenly prevented from entering the northern areas (Stemmerik, 1997). It is possible, and even likely, that the demise of Gondwana glaciers and the following buildup of northern sea ice led to a reversal of thermohaline circulation on a global scale. Without knowing the ultimate cause, we suggest that

the temporal coincidence of a global maximum flooding event, termination of cyclothem sedimentation, demise of Gondwana ice sheets, formation of northern sea ice, influx of cool northern oceanic waters along northwest Pangea, and the onset of the PCE constitutes a chain of events that reflects a profound and irreversible palaeoceanographic shift at the Sakmarian–Artinskian boundary (Fig. 7A, B).

3.5. *End-Permian global warming*

Evidence is mounting that the end-Permian environmental turnover was accompanied by significantly warmer climatic conditions throughout the world. Based on his study of palaeosols in Australia and other areas of Gondwana, Retallack (1999) argued for a major greenhouse warming that accompanied, and perhaps followed, a massive turnover in the land plant populations of the southern hemisphere. He believes that the climate shifted from relatively cold and humid during the Late Permian as expected from the high palaeolatitudes of the areas he studied, to much warmer and dryer during the Early Triassic unexpected for the palaeolatitudes of the area. In that vein, proponents of the idea that a major anoxic event was associated with the end-Permian extinction have more than once suggested that it is a warmer global environment that led to sluggish circulation and fostered anoxia (Wignall and Hallam, 1992). The brief return to a carbonate mud-forming environment in the transitional unit at the base of the Blind Fiord Formation and the presence of redbeds and caliche-bearing fluvial overbank strata in the correlative Bjorne Formation may also provide some evidence for a warmer climatic setting (Fig. 1).

The end-Permian global warming provides a simple explanation for the collapse of chert factories along northwest Pangea. Indeed, warming would have led to a net reduction, and perhaps complete thawing, of northern sea ice hence eliminating the engine of thermohaline circulation (Fig. 7E). This would have led to the breakdown of the conditions favourable to the production, accumulation and preservation of biogenic chert. A much warmer marine environment accompa-

nied by sluggish circulation of oxygen-, silica- and nutrients-poor waters would have applied significant stress upon biogenic productivity, while increasing the solubility of biogenic silica in the water, on the sea-floor and in the shallow burial environment. Likewise, warmer conditions would have favoured the inorganic production of carbonate mud diluting any biogenic silica that may have been produced. The sluggish circulation would also have favoured pervasive oceanic stratification and life-stressing sea-floor anoxia (Wignall and Hallam, 1992; Wignall and Twitchett, 1996; Wignall et al., 1998).

The end of the PCE and the ensuing ETCG do not, however, correspond to a simple return to pre-PCE conditions. The absence of Early Triassic chert worldwide attests for a Mesozoic world that is much different than the Palaeozoic world it replaced. In this case, the severity of the shift may call for both a collapse in productivity of silica-secreting organisms and a set of conditions adverse to the preservation of bedded chert, perhaps including a low supply of dissolved silica to the oceans. As pointed by Kump et al. (2000), although the accumulation of biogenic silica is not a direct indicator of silica weathering intensity, it is a probable indicator of silica supply to the oceans. It is possible that one, of many, contributing factors to the ETCG may have been the uniformly warm and dry mega-monsoonal climate that prevailed throughout Pangea's interior in the Early Triassic (Parrish, 1993), which coupled with the lower topography of the now eroded Hercynian and Ural mountains, may have led to significantly less weathering than during the Permian.

Racki (1999) was puzzled over the absence of chert following the end-Permian extinction event especially since an increase in chert accumulation appears to have accompanied other extinction events such as the Cretaceous–Tertiary and the Frasnian–Fammenian events. As an alternate explanation, he offered the idea that the ETCG reflects the non-selective loss of silica-secreting fauna associated with the end-Permian extinction. Accordingly, Kozur (1998a,b) documented a substantial decline in the number of radiolarian families (nine to two), Demospongiae families (12 to five) and Hexactinellida families (six to one). The

extinction of several families of siliceous sponges and radiolarians is undoubtedly significant and must have contributed to the ETCG as suggested by Racki (1999). However, while the diversity of siliceous organisms remained very low for most of the Early Triassic, these organisms staged a spectacular recovery starting in the latest Early Triassic and climaxing during the early Middle Triassic (Anisian). Interestingly, more than 50% of Middle Triassic siliceous sponge taxa are so-called Lazarus taxa that existed prior to the Late Permian decline (Kozur, 1998a,b). It appears that the negative environmental conditions that led to the demise of the PCE, and surely contributed to the extinction of siliceous biota, prevailed for millions of years, during which many silica-secreting organisms survived and reproduced in yet unknown refuges, patiently awaiting the return of a more hospitable marine environment. It took a staggering 8–10 Ma for the world oceans to fully recover from the end-Permian event.

4. Conclusions

A significant global event occurred at about the Sakmarian–Artinskian boundary that triggered a vigorous pattern of thermohaline circulation along the northwest margin of Pangea. This event followed the thawing of Gondwana ice sheets in the southern hemisphere and coincides with maximum flooding on global scale, cessation of cyclothem sedimentation and onset of widespread biogenic silica preservation (Permian Chert Event). Seasonal thawing of northern sea ice, the presumed engine of thermohaline circulation, led to the influx and circulation of dense, hypersaline cold water in the deep Panthalassan Ocean parallel to the northwest margin of Pangea, displacing nutrients- (phosphorus, iron, nitrogen, etc.) and silica-rich water in its path and forcing it along the oceanic margins via upwelling. Lighter melt water was also carried southward all along northwest Pangea through Coriolis forcing. This led to the contraction and ultimate disappearance of prolific warm-water carbonate factories and their replacement by biologically impoverished cool- to cold-water carbonate factories. This also led to

the expansion of siliceous factories and preservation of their sedimentary products as bedded spiculitic chert, associated with unusually large phosphate and glauconite deposits. The areas favourable to siliceous sedimentation and preservation expanded both landward and southward during the Guadalupian and this process reached its zenith during the Lopingian, at which time not only northwest Pangea but many warmer areas of Panthalassa and the Tethys became the loci of sponge and radiolarian siliceous sedimentation. At that time, seasonal sea ice may have reached the latitudes of the Barents Sea and the Sverdrup Basin.

A major event occurred during latest Permian time which led to the collapse of thermohaline circulation. Presumably rapid global warming and equally rapid thawing of northern sea ice destroyed a palaeoceanographic setting that had promoted the growth and preservation of biogenic siliceous sediments for nearly 30 Ma. Complete thawing of northern sea ice would have ended thermohaline circulation and led to warm and sluggish oceanic conditions inimical to the production, accumulation and preservation of biogenic silica. These conditions lasted between 8–10 million yr (Early Triassic Chert Gap), during which several families of silica-secreting organisms became extinct, while others took refuge under more favourable conditions, before staging a come back during Anisian time (early Middle Triassic).

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