Chapter 2
Geological Framework

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2.1 Plate Tectonic Evolution of the Arctic

The Arctic Ocean comprises two main deep subocean basins, the Amerasia and Eurasia basins, separated by the elongate Lomonosov Ridge (Fig. 2.1). In a plate tectonic framework the Eurasia Basin is linked to the Norwegian-Greenland Sea and the Atlantic through the northernmost part of the Eurasia-North America plate boundary (Fig. 2.2). The plate boundary comprises two mid-ocean ridges, the Knipovich Ridge in the Norwegian-Greenland Sea and the Gakkel Ridge in the Eurasia Basin, linked by transform faults and oblique spreading segments in the Fram Strait (Engen et al. 2008). The passive margins flanking the plate boundary feature two marginal plateaus, the Morris Jesup Rise and the Yermak Plateau.

The Norwegian-Greenland Sea and the Eurasia Basin developed in a three-plate setting as Cretaceous-Paleocene rifts propagated northwards on either side of Greenland (e.g., Talwani and Eldholm 1972). By earliest Eocene time, seafloor spreading was established in both basins (Talwani and Eldholm 1972; Vogt et al. 1979), while relative motions between the Eurasia and Greenland plates were accommodated by a megashear region, the De Geer Zone, in the emergent ocean basin between Svalbard and Greenland (Harland 1969; Faleide et al. 1993, 2008). At the Eocene-Oligocene transition, Greenland became part of the North American plate and transform faulting in the Greenland Sea was replaced by oblique rifting and seafloor speading (Talwani and Eldholm 1972).

The opening of the Eurasia Basin split off a segment of the Eurasian continental margin that subsequently became the Lomonosov Ridge continental sliver (e.g., Grantz et al. 2001). The inner part of the Yermak Plateau is continental while its outer part and the Morris Jesup Rise have been attributed to voluminous volcanism at the Eurasia-Greenland-North America extinct triple junction (e.g., Jackson et al. 1984). Continental remnants in the Norwegian-Greenland Sea include the

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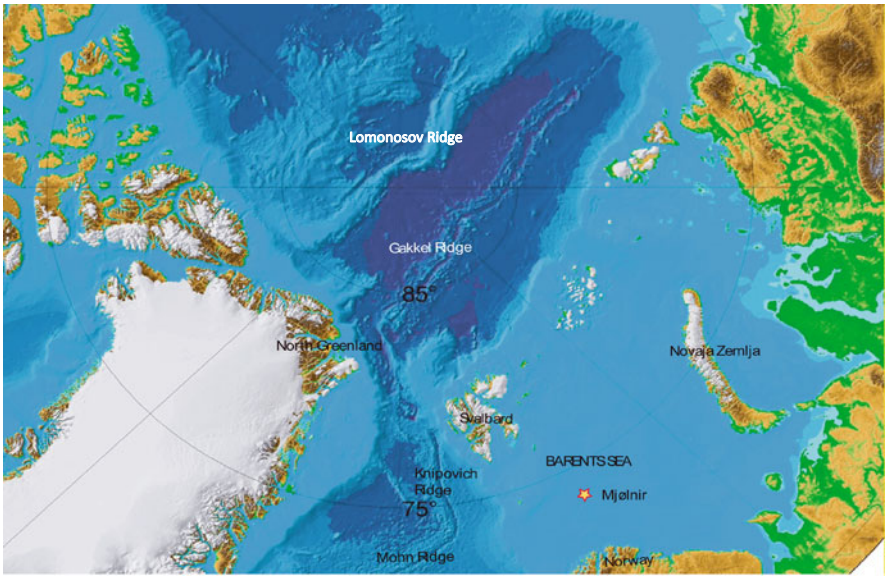

Fig. 2.1 Map of the Arctic, showing the location of the Mjølnir impact (*star*) between Svalbard and the mainland Norway. Modified from Jakobsson et al. (2003)

Hovgård Ridge microcontinent (Myhre et al. 1982) and, possibly, the Greenland Ridge (Tsikalas et al. 2002b; Døssing et al. 2008).

2.2 Mesozoic Stratigraphy and Depositional Environments of the Arctic

The Barents Shelf consists of several kilometers thick Paleozoic to Cenozoic sedimentary successions. From seismic lines crossing the Mjølnir crater it is evident that the impact disturbed strata down to, but not including the Upper Paleozoic strata (Figs. 1.4 and 1.10).

The Upper Palaeozoic succession (Larssen et al. 2005; Worsley 2008) comprises basal coarse clastics grading into Carboniferous carbonates and evaporites. Carbonate deposition with thick reef complexes continued into the Permian. Towards the Late Permian the northward drift of the land areas and the fusing of Laurasia and Siberia by closing the Uralian Sea, sealed the contact with the southern latitude seas. On this northern margin of the thereby formed Pangea supercontinent the northern ocean embayment changed from shallow to deep shelf setting within a cool climate. The abundant silica sponges resulted in cherty sediments forming the uppermost Permian succession. The top of this unit is visible as a pronounced seismic reflector throughout the Barents Shelf, and is seen as a marked ledge in the landscape on Svalbard. The reflector does not appear to be disturbed by the Mjølnir impact in the target area (see Chaps. 3 and 4).

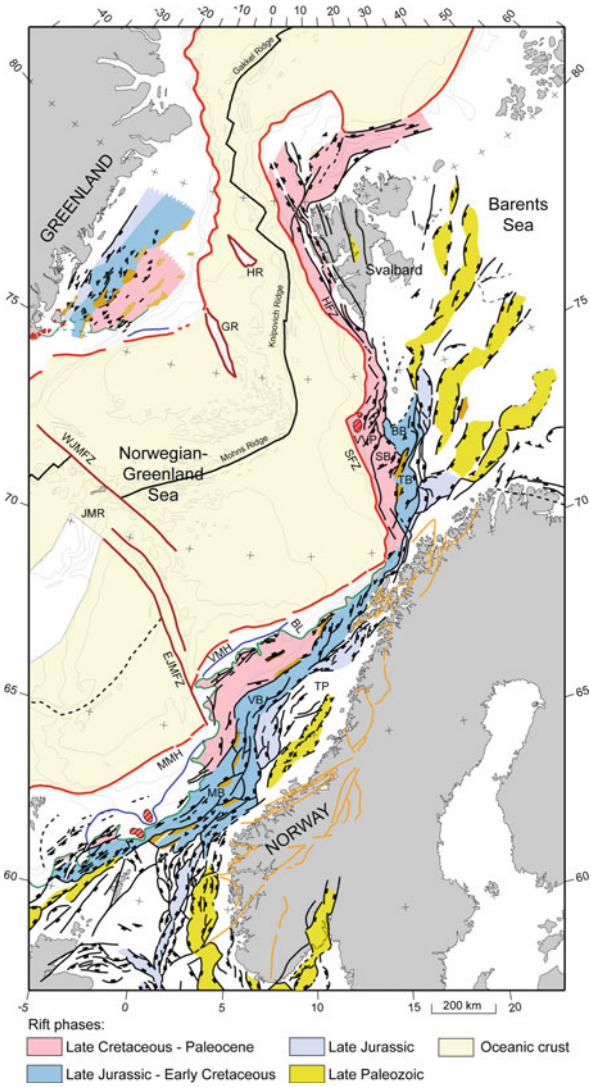

Fig. 2.2 Tectonic map of the North Atlantic and the Barents Sea. The Mjølnir impact crater is located on the Bjarmeland Platform in the Barents Sea. The map outlines the Jan Mayen Fracture Zone in *dark red* (EJMFZ, WJMFZ) and the North Atlantic continental margin (HFZ, BFZ, MMH). Modified from Faleide et al. (2008)

This Palaeozoic-Mesozoic boundary reflects the marked change from the silica cemented cherts to overlying clastic Triassic shales. The Lower and Middle Triassic succession, the Sassendalen Group is shale dominated, but also contains several coarsening upward sequences ending with sandstones landwards and towards the upper regressive part of each sequence (Mørk et al. 1999; Worsley 2008). Palynomorphs and re-deposited sediments from this group are found in the Mjølnir crater infill.

The Upper Triassic and Lower Jurassic succession (Kapp Toscana Group) is sandstone dominated. It represents shallow shelf to deltaic depositional environments (Worsley 2008; Riis et al. 2008). Re-deposited sandstone fragments from this group are abundant in the Mjølnir crater core.

2.2.1 Geological and Palaeogeographical Setting

2.2.1.1 Cretaceous Palaeogeographic Setting

A 142 million year plate reconstruction shows that the Barents Sea region was centered on the 55°N parallel (Fig. 2.3). The impact palaeolatitude is calculated to 56.4°N, at a time when Greenland also bordered and defined the western margin of the Barents Sea. Thus, and with relevance to tsunami modelling of the impact (compare Chap. 10), the distance to Greenland (ca. 300 km) was approximately the same as to Northern Norway (Finnmark), where waves as high as 100 m have been estimated (Glimsdal et al. 2007).

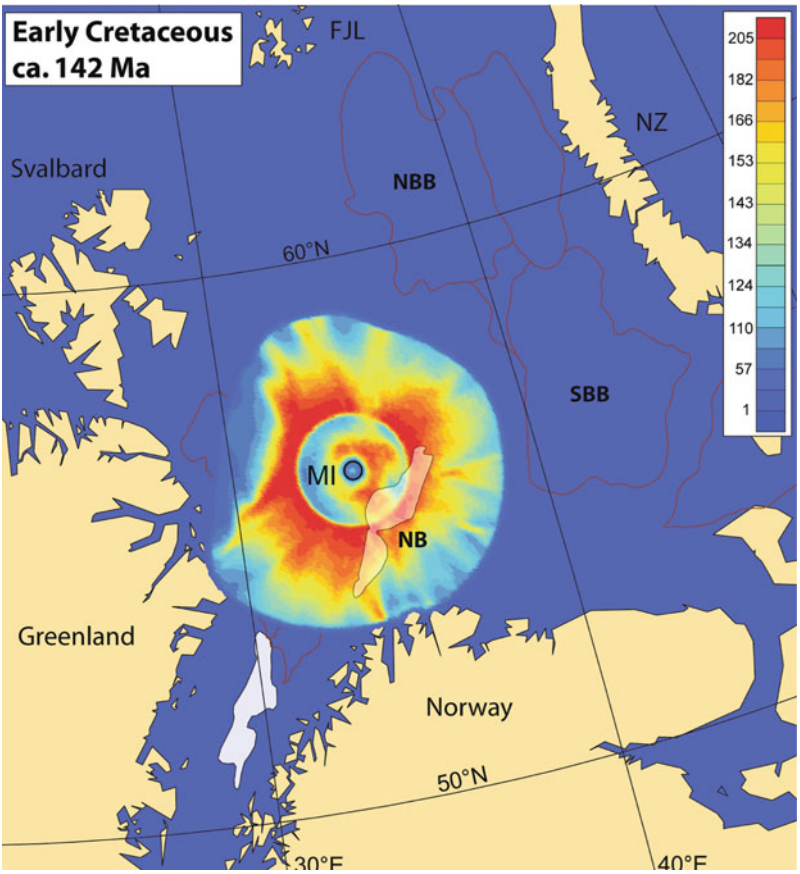

Fig. 2.3 The palaeogeographic setting of the Mjølnir impact site reconstructed to the Lower Cretaceous, at ca.142 Ma. The main differences with respect to the current setting are the palaeolatitude of 56.4°N and the then young and narrow Atlantic Ocean. The estimated tsunami wave height (after Glimsdal et al. 2007) and sedimentary basin outlines are draped on the reconstruction. FJL = Franz Josef Land, NZ = Novaya Zemlya, NBB = North Barents Basin, SBB = South Barents Basin, NB = Nordkapp basin, MI = Mjølnir impact crater

2.2.1.2 The Barents Sea in Time and Space

The Barents Sea realm has drifted northward over the past 300 million years, crossing tropic and sub-tropic latitudes as confirmed in the sedimentary record (Fig. 2.4). Fauna and sedimentary facies are sensitive to climate changes and to latitude. Distribution of evaporites, coal and certain carbonates are strongly latitudinally dependent. Evaporites are mostly deposited in arid sub-tropical regions, whilst coal is formed in wet equatorial regions or the northerly and southerly wet belts. This is well exemplified for the Barents Sea realm: 300 million years ago the Barents Sea

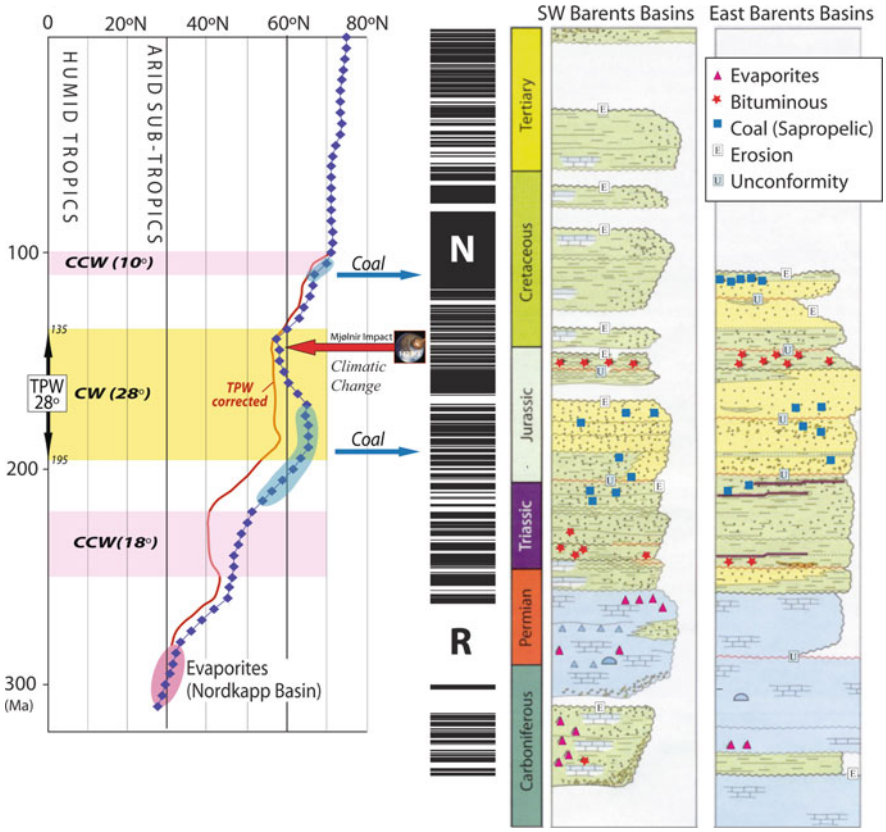

Fig. 2.4 The Barents Sea in space and time. Along with the stratigraphic column for the southwestern Barents Sea and the eastern Barents Sea basins, the magnetic polarity, the latitudinal drift of the Mjølnir impact site is shown. The drift curve is derived from a global apparent polar wander path with (*red curve*) and without (*blue curve*) corrections for true polar wander (TPW; Torsvik et al. 2008, Steinberger and Torsvik 2008). The TPW corrected curve shows a general northward movement of the Barents Sea realm (with respect to the mantle) but the flat Jurassic-Early Cretaceous section of the curve show that the apparent Mid-Late Jurassic southward movement (with respect to the spin-axis) is an artefact of TPW. The episodes of true polar wander are marked in *yellow* for clockwise (CW) and *reddish* for counterclockwise (CCW) motion

was at sub-tropical latitudes, a palaeolatitude that fits well with the occurrences of Late Carboniferous-Early Permian evaporites in the Nordkapp Basin. Subsequently the Barents Sea drifted northward, and by late Triassic to Jurassic time, we find coal witness that the region had entered the northern wet belt. We also show that the Barents Sea region drifted somewhat southward during Mid-Late Jurassic before continued northward drift (incidentally coinciding with the Mjølnir impact event) with coal once again appearing in the East Barents deposits (Fig. 2.4). These back-and-forth N–S movements (ca. five degrees in total) led to a climatic change, not because of "continental drift" but due to true polar wander (TPW). TPW is the rotation of the entire solid Earth's outer shell with respect to the spin-axis, and during Jurassic and Early Cretaceous time (195–135 Ma) the entire Earth rotated ca. 28° clockwise (Steinberger and Torsvik 2008). This lead to slow climatic changes with some areas becoming warmer, such as the Barents Sea moving away from the spin axis, whilst others became colder.

2.2.2 Svalbard

The Mesozoic deposits on Svalbard form a 3-km-thick succession of siliciclastic sedimentary rocks (Figs. 1.3, 1.4, 2.5 and 2.6). On Svalbard the Mesozoic succession displays repeated sequences of nearshore to shelf deposits, dominated by coastal progradation and deltaic sediments. The Mesozoic sediments represent

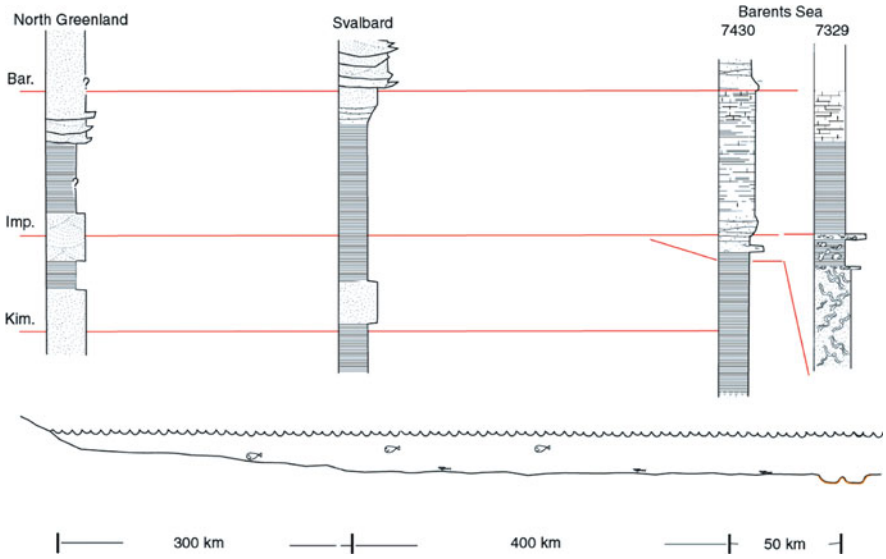

Fig. 2.5 Generalized stratigraphic correlations between Barents Sea, Svalbard and North Greenland close to the time of the Mjølnir impact. Bar. = Barremian, Imp. = Time of impact, Kim. = Kimmeridgian

Fig. 2.6 Geological map of Svalbard (Dallmann et al. 1999)

stable platform depositional conditions, dominated by fine-grained sedimentation, shales, silt-, and sandstones.

The Triassic Sassendalen Group (Fig. 1.4) (Mørk et al. 1982, 1999) comprises coastal and shallow marine sediments in the west (i.e., Vardebukta, Tvillingodden and Bravaisberget formations) grading eastwards into open marine shale dominated shelf sediments (i.e., Vikinghøgda and Botneheia formations).

The succeeding Kapp Toscana Group (Mørk et al. 1982, 1999) displays a shift in sediment source areas from W and SW to SE. Extensive deltas filled the basin from southeast, Tschermakfjellet Formation representing the base with prodeltaic shales. This unit is overlain by De Geerdalen Formation with extensive deltas transporting sediment from mainland Norway and from the Urals filling most of the Barents Sea and Svalbard with deltaic to shallow shelf sediments (Mørk et al. 1982; Riis et al. 2008).

The lithological relations vary both qualitatively and quantitatively throughout the Jurassic and Cretaceous. The Janusfjellet Subgroup (thickness variations; 280–750 m) generally spans the time interval from Middle Bathonian to Barremian (Figs. 1.4 and 2.5). The Janusfjellet Subgroup demonstrates an overall, wide, continuous sedimentation, interrupted by stratigraphic breaks in the Middle Oxfordian and Upper Volgian (Smelror 1994). Its stratigraphical and sedimentological development is well known from the area, forming parts of the foundation of our study of the Mjølnir impact and the regional correlation of the Mjølnir event (Parker 1967; Birkenmajer 1980; Dypvik 1980; Nagy et al. 1990; Dypvik et al. 1991a, b; Nagy and Basov 1998)

The Janusfjellet Subgroup makes up the lower part of the Adventdalen Group and can be correlated with the Fuglen/Hekkingen/Klippfisk/Knurr formations in the Barents Sea (Mørk et al. 1999) (Fig. 1.3). The shales and sandstones of the Janusfjellet Subgroup comprise the Upper Jurassic Agardhfjellet Formation and Lower Cretaceous Rurikfjellet Formation (Figs. 1.3, 1.4 and 2.7).

The Agardhfjellet Formation is 242 m in thickness in the stratotype and consists of dark shales and claystones with a few sandier units dispersed in the lower (Oppdalen Member) and middle parts (Oppdalsåta Member) (Figs. 1.3 and 1.4). Typically also highly organic rich, papery shales (paper shales), often with more than 10 wt% TOC, are found in the black to dark grey shale units of Lardyfjellet and Slottsmøya members (Fig. 2.8).

The lowermost sandy unit, Oppdalen Member (10–60 m), is characterized by a fining upwards development from poorly sorted, silty sandstones into dark grey, silty shales. These shallow marine shales develop further into the organic rich, paper shales of the Lardyfjellet Member, which were deposited during anoxic to dysoxic conditions (Figs. 1.3 and 2.8) (Dypvik et al. 1991a). The Oppdalen and Lardyfjellet members represent the Middle Bathonian to Oxfordian units in the area. Succeeding these shales follows the bioturbated sandstones of the overlying Oppdalsåta Member. These sandstones and sandy shales are, on the average 28 m in thickness, mainly forming coarsening upwards successions of Oxfordian to Kimmeridgian age. The Oppdalsåta Member has been interpreted by Dypvik et al. (1991b) to represent storm deposits, formed in the wide, epicontinental

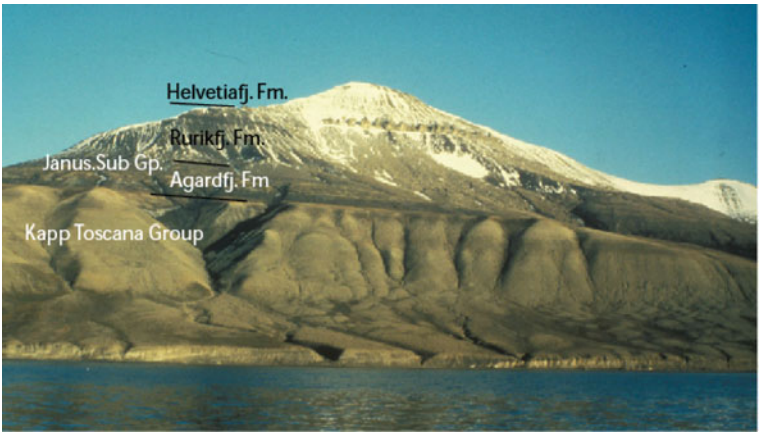

Fig. 2.7 The Janusfjellet Subgroup at Wimanfjellet, Svalbard. The Janusfjellet Subgroup, Agardhfjellet Formation, Rurikfjellet Formation, and Helvetiafjellet Formation are marked (Photo: Jenö Nagy)

Fig. 2.8 Close up (photo about 30 cm in height) photos of paper shales from Agardhfjellet Formation, Wimanfjellet

paleo-Barents Sea. The vivid sea floor life returned to the region immediately after the stormy period of the Oppdalsåta Member and disturbing these sandy formations completely. Consequently, the coarsening upwards Oppdalsåta sandstones are characterized by homogeneous appearance, due to severe post-storm bioturbation. The Oppdalsåta Member is today rich in parallel and current oriented belemnites, but poor in sedimentary structures other than the signs of thorough bioturbation.

The uppermost part of the Jurassic succession is made up of the dark grey to black, organic rich shales of the Slottsmøya Member. It often reaches more than

100 m in thickness and consists dominantly of paper shales, with varying degrees of fissile ability.

In central Spitsbergen the Agardhfjellet Formation is succeeded by the Myklegardfjellet Bed (mainly Ryazanian in age), which is 0–11 m in thickness, and forms the base of the succeeding Rurikfjellet Formation. The Myklegardfjellet Beds consists of soft, often yellowish to greenish plastic clays commonly rich in dolomite, pyrite and sporadic altered glauconites present in separate layers. The Myklegardfjellet Bed deposits may represent the transgressive start of a flooding event, marking the transition from shallow shelf to relatively deep mid-shelf conditions in the succeeding Wimanfjellet Member. The depositional development continues into the regressive prodeltaic deposits of the uppermost part of the Rurikfjellet Formation (Figs. 1.3 and 1.4) (Dypvik et al. 1991a).

The Rurikfjellet Formation (Wimanfjellet and Ullaberget members) often reaches more than 200 m in thickness and is composed of dark grey, commonly silty shales. Increasing amounts of siltstone and sandstone occur in its upper part, commonly in upwards coarsening units. The lowermost part of the Rurikfjellet Formation, the Myklegardfjellet Bed (Fig. 2.9), is succeeded by the very fine claystones of the Wimanfjellet Member. They form the finest grained unit of the

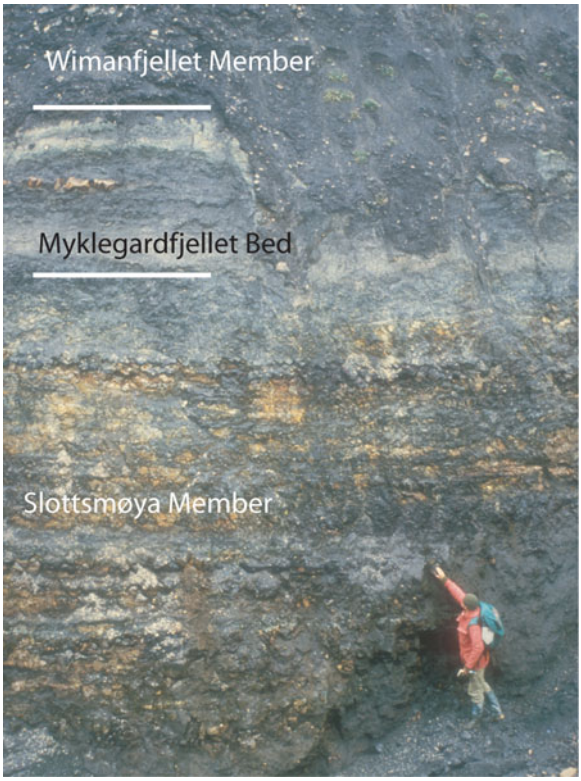

Fig. 2.9 The Myklegardfjellet Bed is located in between the Slottsmøya Member (*below*) and the Wimanfjellet Member above at Glitrefjellet (Reindalen, Svalbard). The possible Mjølnir impact level is located at person's hand

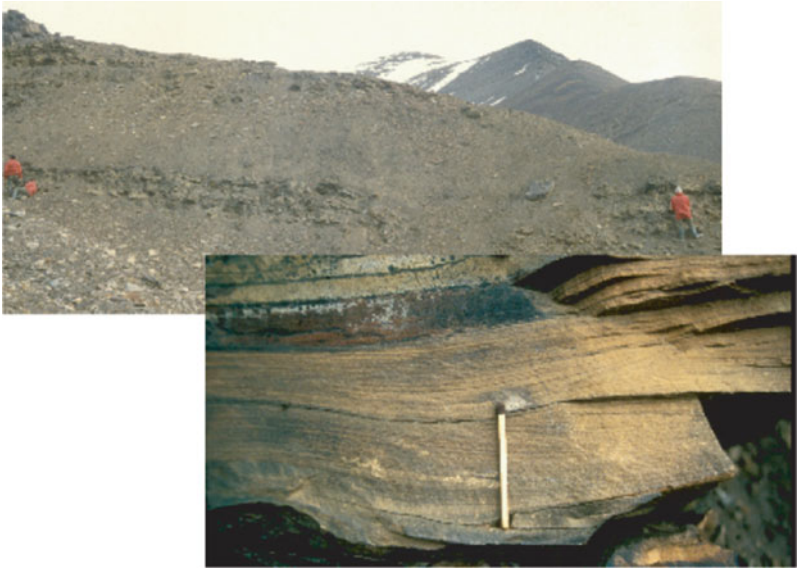

Fig. 2.10 The Ullaberget Member of the Rurikfjellet Formation at Aasgaardfjellet, with a few sandstone beds is seen in the upper photo (persons for scale). The lower photo is a close up of a hummocky cross stratification unit (HCS) found in these beds

Janusfjellet Subgroup, representing a relative deep shelf setting at maximum flooding stage. The smectitic claystones of the Wimanfjellet Member are succeeded by the general coarsening upwards successions of the Ullaberget Member, 155 m in thickness in the stratotype (Fig. 2.10). In contrast to the mid-shelf facies of the Wimanfjellet Member the Ullaberget Member represents prodeltaic depositional environments, sedimentologically related to the succeeding deltaic deposits of the Helvetiafjellet Formation.

The Barremian Helvetiafjellet Formation varies from 40 to 155 m in thickness (Figs. 1.3 and 2.11). It consists in the lower part of the coarse sandstones of the pronounced Festningen Member (up to 16 m in thickness), which is covered by coal-bearing successions. The Helvetiafjellet Formation represents complex transgressive interacting fluvial, delta plain, mouth bar, barrier bar, tidal estuary and transgressive sheet sandstone facies succeeding a period of relative sea level fall (Gjelberg and Steel 1995; Mørk et al. 1999; Midtkandal et al. 2007). The Festningen Member is a fluvial dominated delta complex prograding from the northwest. It correlates with the lower part of the Kolmule Formation of the Barents Sea. The Helvetiafjellet Formation displays an overall transgressive trend (Gjelberg and Steel 1995), but contains numerous punctuated regressive pulses within its general transgressive development.

The Aptian–Albian Carolinefjellet Formation has large thickness variations, from 190 m in the northern areas to more than 1,200 m in the southeastern part of Svalbard. The Carolinefjellet Formation consists of alternating shallow marine

Fig. 2.11 The Helvetiafjellet Formation, the Festningen Sandstone Member as it appears at Lardyfjellet in the overview photo to the *left* (sandstone unit about 6 m high). Well-developed cross-bedding in the sandstones of the same locality is shown in the *right* photo

shales and sandstones, reflecting prodeltaic to distal marine depositional conditions. It correlates with the upper part of the Kolmule Formation in the Barents Sea (Fig. 1.3).

The Cenozoic beds of Svalbard span the time interval from Late Paleocene to Oligocene, and represent cyclic deposition in a foreland depression (Dallmann et al. 1999). The succession consists of intermixed continental and marine clastics. A thick pile of more than 2,500 m of Cenozoic sediments covered large parts of the central and southern Svalbard. Maturation studies of organic matter indicate that an even 1,000 m thicker original Tertiary succession has been present in Svalbard (Manum and Throndsen 1978).

2.2.3 Barents Sea

In the Barents Sea the Lower Triassic sediments show moderate deep-shelf facies consisting of mixed shales and sandstones assigned to the Havert Formation (Worsley et al. 1988) (Fig. 1.4). It is correlated with the Vardebukta Formation and lower part of the Vikinghøgda formations on Svalbard (Mørk et al. 1999). This unit is succeeded by the sandstone, siltstone and shales of the Klappmyss Formation (Worsley et al. 1988) correlated to the Tvillingodden Formation and upper part of Vikinghøgda Formation on Svalbard (Mørk et al. 1999).

The Middle Triassic succession comprises claystones and sandstones of the Kobbe Formation (Worsley et al. 1988), however in central part of the Barents Shelf, it is replaced with the organic-rich Steinkobbe Formation (Mørk and Elvebakk 1999). The Kobbe Formation dominates on the eastern part of the basin and is an equivalent of the Bravaisberget Formation on western Svalbard. The time equivalent Steinkobbe Formation is an equivalent of the Botneheia Formation of central and eastern part of Svalbard (Mørk et al. 1999).

The Late Triassic uplift of the area north of Svalbard and the Norwegian–Kola land areas resulted in dominantly sand deposition in the paleo-Barents Sea. The

Snadd Formation mainly got its sediment supply from the south and southeast and form extensive delta front to delta top deposits, equivalent to the De Geerdalen Formation of Svalbard (Worsley 2008; Riis et al. 2008). Both on Svalbard and in the Barents Sea the uppermost Triassic and Lower/Middle Jurassic succession consists of extensively reworked sandstones; i.e. the Wilhelmøya (on Svalbard) and Realgrunnen (in the Barents Sea) subgroups, respectively. In the Barents Sea, it is formed by the deltaic to shallow marine Fruholmen Formation and the overlying shallow marine to coastal Stø Formation (Worsley et al. 1988; Worsley 2008).

The Mesozoic sediments of the Barents Sea, the Adventdalen Group in particular, can be recognized and match with similar deposits in Svalbard (Figs. 1.3, 2.5 and 2.6). These correlations are as follow; the Fuglen Formation correspond to the Oppdalen and Lardyfjellet members of the Agardhfjellet Formation, the Hekkingen Formation can be correlated to the Oppdalsåta and Slottsmøya members, while Klippfisk and Knurr formations are equivalent to the lowermost part of the Rurikfjellet formation (Wimanfjellet Member). The Kolje Formation can be correlated to the Ullaberget Member of Svalbard. The Kolmule Formation of the Barents Sea can be match the Helvetiafjellet and Carolinefjellet formations of Svalbard (Figs. 1.3, 1.4 and 2.5).

In the southeastern Barents Sea the Fuglen Formation varies between 20 and 50 m in thickness, but it reaches more than 200 m in the Troms III area (Worsley et al. 1988). The formation is of Late Bathonian to Middle Oxfordian age. It consists of mudstones and minor limestones and represents marine shelf deposits.

In the Barents Sea the Hekkingen Formation, which is of Late Oxfordian to Ryazanian age, has a thickness varying between 110 and 400 m. It consists of mostly dark grey to black shales, with some silt and sandstone beds dispersed. The Alge and Krill members form the lower and upper parts of the Hekkingen Formation, respectively. Both these units are dominated by dark grey, to black organic rich shales, which is reflected in the names of the two members (Figs. 1.3 and 2.12). In the Barents Sea exploration wells the Fuglen Formation is recognized by its characteristic log response, in particular high gamma activities (Dypvik et al. 2004). The highest gamma response is, however, found in the Alge Member of the Hekkingen Formation and correlate with increased uranium concentrations in this organic rich member with TOC values up to 12%.

The Lower Cretaceous Hekkingen Formation is succeeded by the Lower Cretaceous Knurr and Klippfisk formations. The Knurr Formation is a Volgian to Barremian unit, consisting of mudstones, sandstones and limestones. The formation represents distal marine shelf depositional conditions, with only locally restricted bottom ventilation. The Knurr Formation varies from 56 to 285 m in thickness. The time-equivalent Klippfisk Formation comprises condensed carbonate deposits preserved on structural highs and platforms (Fig. 2.13). These carbonate platform deposits have been found as far east as in the Olga Basin and it outcrops on Kong Karls Land on Svalbard (Smelror et al. 1998). The Klippfisk Formation is thin, generally only from 4 to 15 m in thickness and consists of marls and limestones of Berriasian to Hauterivian age.

Fig. 2.12 The Alge member of the Hekkingen Formation from the Barents Sea. IKU (Continental Shelf Institute Norway) core 7430/10-U-01. The finely laminated dark grey shales are seen disrupted by very thin silt laminae and a large pyrite concretion in the upper part of the core

In the Barents Sea the succeeding Kolje Formation, which is between 15 and 403 m in thickness, consists mainly of shales and mudstones representing distal, open marine depositional conditions with moderate water circulation. It forms a transgressive phase on top of the Klippfisk/Knurr formations couplet.

2.2.4 Greenland

In North Greenland, the Mesozoic Wandel Sea Basin comprises a succession with Lower Carboniferous to Paleogene sediments. This succession was deposited on the margin of the stable Greenland craton where the Caledonian and Ellesmerian orogenesis intersected. On top of these metamorphic Precambrian to Silurian formations Upper Paleozoic to Lower Mesozoic sedimentary sequences are found in an extensional setting, with four main tectonic events recognized: one in mid-Jurassic, one in mid-Cretaceous, one at the end of Cretaceous and one post-Paleocene extensional event. This last event took place after the tectonic shift of the plate boundary towards its present position between Greenland and Svalbard (Håkansson et al. 1993). Accumulation of Carboniferous and younger sediments postdate the Caledonian and Ellesmerian Orogeny. In Kilen a more than 3 km-thick succession

Fig. 2.13 The core piece shows bioturbated carbonates from the Klippfisk Formation. It was cut from IKU (Continental Shelf Institute Norway) core 7430/10-U-01

of such sediments are found, including about 1,400 m of Upper Triassic, Jurassic, and Cretaceous deposits (Fig. 1.3).

Jurassic formations are widely distributed in East Greenland, but are only found in two major sites in North Greenland: Kilen and East Peary Land. These North Greenland locations are recognized by coastal and shallow marine clays and well sorted cross-bedded sands (Håkansson et al. 1994; Heinberg and Håkanson 1994; Dypvik et al. 2002). In particular, the Lower Cretaceous sand units are well developed; as seen in the white, well sorted, sandstone formations of the Ladegårdsåen (East Peary Land) and the Lichen Ryg (Kilen) formations (Fig. 1.3). In both areas, the uppermost part of the Jurassic succession consists of dark grey, organic rich shales, commonly containing plant fragments (Fig. 2.14). The successions comprise upward coarsening, 5–30 m thick sequences, covering the Oxfordian to Valanginian time interval.

In the Late Jurassic depositional models of Dypvik et al. (2002), the North Greenland area forms the southwestern coastline of the palaeo-Barents Sea, with a possible opening towards the Canadian Sverdrup Basin (Figs. 2.15 and

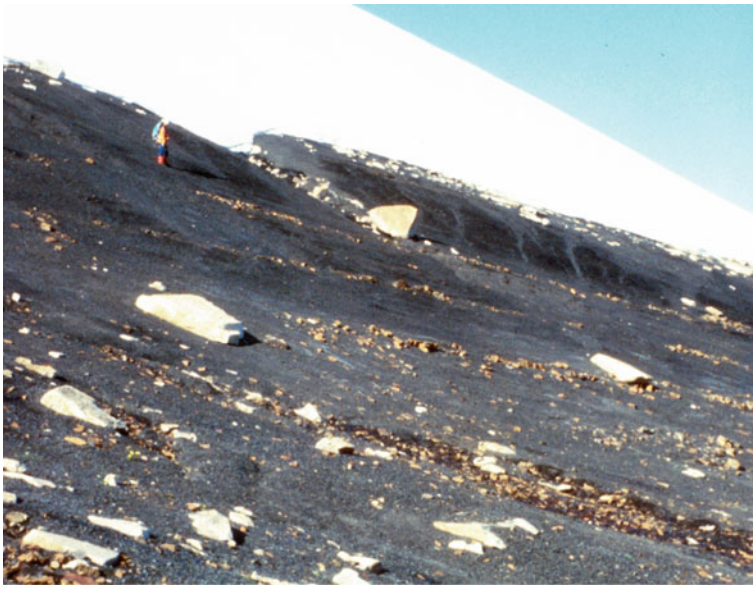

Fig. 2.14 Upper Jurassic dark grey shales of the Dromledome Formation in Kilen, North Greenland. Thin siderite beds are found throughout the formation, see, e.g., one such bed just *below* the person

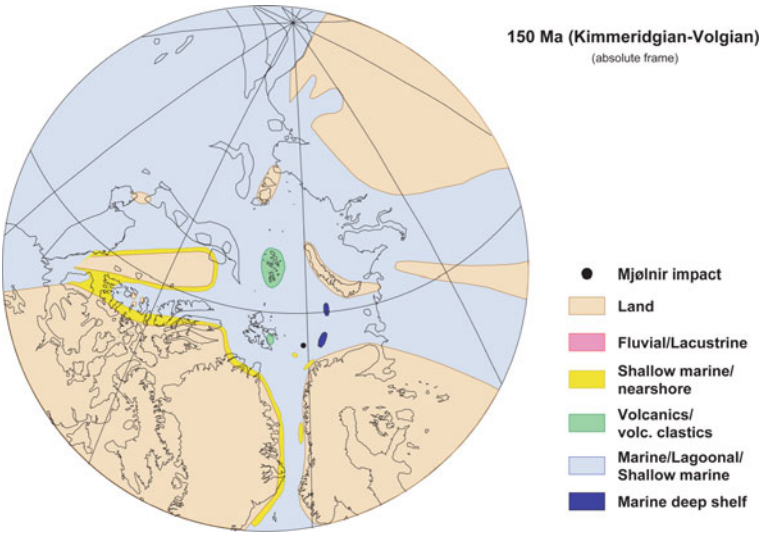

Fig. 2.15 The paleogeographic setting during the Kimmeridigan-Volgian (∼150 Ma) when the Mjølnir bolide hit the paleo-Barents Sea is shown. Based on the plate reconstruction of Lawver et al. (1999)

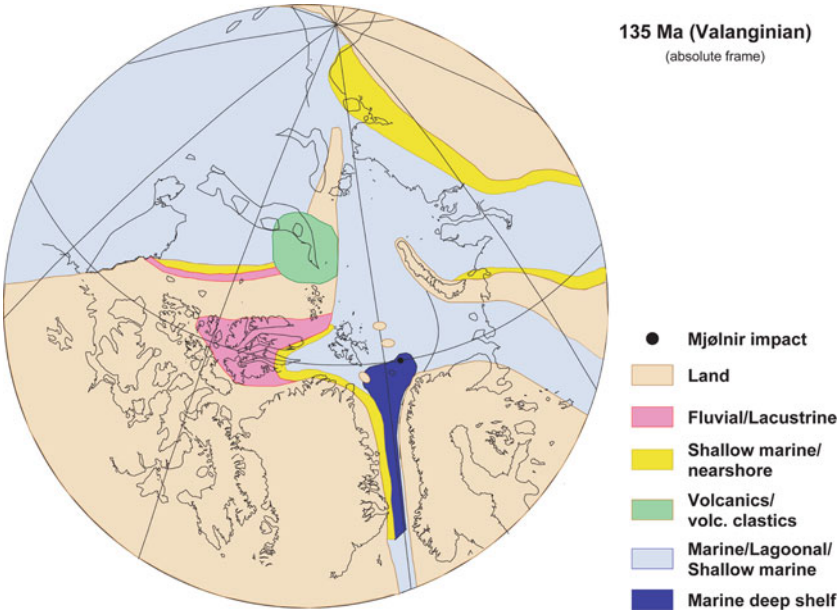

Fig. 2.16 The paleogeographic setting during the Valanginian (~135 Ma) is shown. The possible opening/deep water connection through the pre-North Atlantic are discussed. The reconstruction is based on Lawver et al. (1999)

2.16). This opening was closed in Early Cretaceous time in connection with the northerly uplift/doming of the region, most likely related to breakup in the Amerasia Basin and associated regional magmatism within an Arctic large igneous province (LIP) and formation of the Alpha Ridge (Grogan et al. 1998; Maher 2001). The doming was initiated in Valanginian time and continued through the Early Cretaceous.Valanginian to Hauterivian/Barremian regressive sequences are developed in North Greenland, but comparable units are also evident in the Sverdrup Basin (Canada) (Embry 1991) and Svalbard (Helvetiafjellet Formation).

From the early Eocene (53.3 Ma), sea floor spreading took place along the Gakkel Ridge in the Arctic Ocean (Eurasia Basin) and the Mohns Ridge in the Norwegian–Greenland Sea, accompanied by strike-slip movements between Svalbard and North Greenland (Skogseid et al. 2000; Eldholm et al. 2002; Engen et al. 2008) (Figs. 2.1 and 2.2). The strike-slip movements (and associated opening of the southern Greenland Sea) continued to the Eocene/Oligocene transition, connecting the spreading basins of Arctic Ocean and Norwegian Greenland Sea. At this point in time (33.3 Ma) the movements between Svalbard and NE Greenland shifted towards oblique extension, and the Fram Strait deep-water gateway opened by sea floor spreading in the Miocene (Lawver et al. 1990; Faleide et al. 1993; Eldholm et al. 1994; Torsvik et al. 2001; Engen et al. 2008). Since then, spreading has taken place along this major lineament with final establishment of the present seafloor spreading regime at about 10 Ma.

2.2.5 Siberia

The Triassic succession of Siberia is partly comparable to the time-equivalent Barents Sea succession (Egorov and Mørk 2000). During the Late Jurassic and Early Cretaceous the Siberian region formed a stable biogeographical ecotone with faunistic changes to a large extent controlled by changes in sea-level (Zakharov and Rogov 2003). The Siberian Arctic was a major part of the wide, epicontinental paleo-Arctic Sea, which consisted of three branches opening towards the south (Zakharov et al. 2002). These depositional basins are named, from west to east; the Pechora, West Siberian and Khatanga/Anabar basins. The Ural Mountains separated the Pechora and West Siberian basin, which in turn was separated from the Khatanga/Anabar Basin by the Central Siberian landmass. In the Pechora Basin the Jurassic-Cretaceous succession reaches a composite thickness of about 370 m (120 m Cretaceous and 250 m Jurassic) (Malinovsky et al. 1999; Dypvik and Zakharov 2010).

The sedimentation in the different Siberian basins during the Jurassic and Early Cretaceous took place in a wide range of environments from proximal alluvial, via lacustrine swamps, through shallow marine in Early and Middle Jurassic, and dominating open marine in Late Jurassic and Early Cretaceous. The Upper Jurassic, mainly marine clays, were deposited in the central regions of the basin, and are presently found as black and grey organic rich shale formations (Figs. 2.17 and 2.18). Anoxic bottom water conditions dominated in wide central regions, while coarser grained, sand sedimentation took place along the more ventilated margins of the basins. Finely laminated organic-rich, black shales, without any visible bioturbation, dominated in the basin central areas of the Siberian basins (Zakharov and Rogov 2003). Marginally, carbonate cemented sandstones are commonly found. These are often enriched in phosphates and glauconite. Generally, the amount of marine indicators increases from south to north and towards the opening to the paleo-Arctic seas (Malinovsky et al. 1999).

The prolific hydrocarbon source rock of the Bazhenov Formation in the West Siberian Basin (Volgian to Berriasian in age) is made up of black to brown, organic rich shales (Gavshin and Zakharov 1996; Zakharov et al. 1998). It was deposited during a period of 5–6 million years and it is on the average about 25–30 m in thickness. The formation represents a basinal depositional facies, covers more than 1 million km^2, and is normally buried beneath 2,000–3,000 m of younger sediments. It contains on the average weight 8% TOC and can be correlated to the Hekkingen Formation of the Barents Sea (Fig. 2.12). The Bazhenov shales contain less than 5 weight % of sand and silt, but the formation displays high concentrations of biogenic silica (Zakharov et al. 1998). In the northern, more ventilated environments of the Siberian basins towards the wide Arctic basin, several fossil species similar to the ones discovered in Greenland and Svalbard have been recognized (Zakharov 2004, personal communication).

Marine shelf sedimentation continued into the Early Cretaceous, but at that time the subsidence was reduced and consequently the basins were gradually filled up. This regressive trend dominated in the Early Cretaceous, but was interrupted by

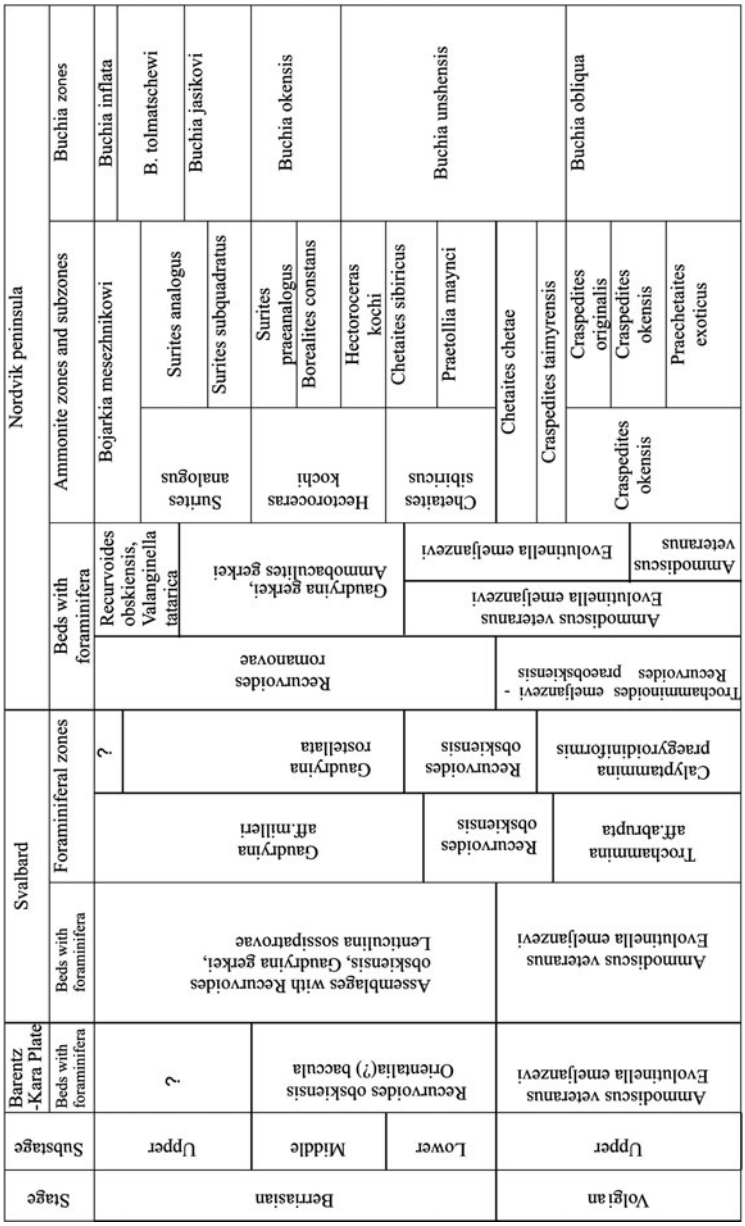

Fig. 2.17 The figure shows a comparison of Siberian and Barents Sea Volgian and Berriasian detailed biostratigraphy. It is based on Dypvik and Zakharov (2010)

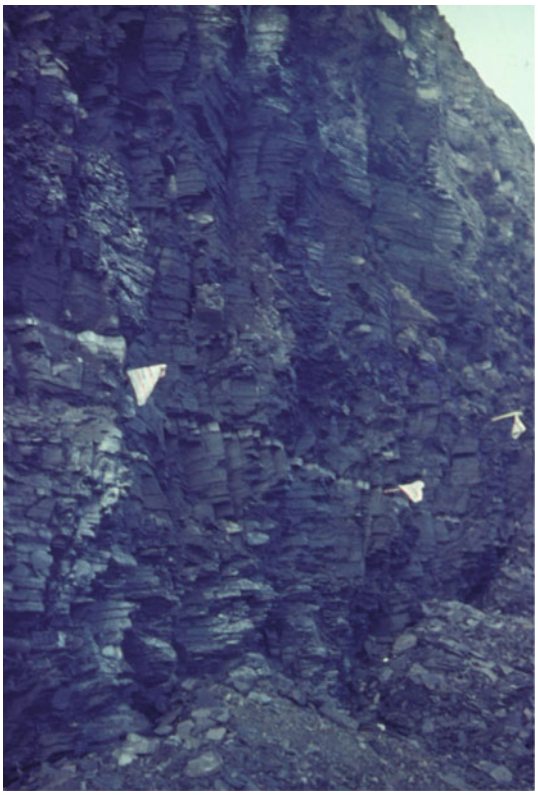

Fig. 2.18 Black shales exposed on the Nordvik Peninsula, West Siberia. The *white flags* showing a phosphate concretionary layer near the Jurassic/Cretaceous boundary. (Photo Victor Zakharov)

several minor transgressive episodes. In the south, deposition of anoxic, organic rich clays still took place in Early Cretaceous time, but increased ventilation in the water masses terminated the organic rich clay deposition.

2.2.6 Late Jurassic and Early Cretaceous Depositional Configuration

The Late Jurassic (Kimmeridgian – Volgian) and Early Cretaceous (Valanginian) paleogeographic reconstructions (Figs. 2.15 and 2.16) of Barents Sea, Svalbard, North Greenland and Siberia are based on our own field studies and detailed literature surveys. The plate tectonic reconstructions are founded on Lawver et al. (1999).

During the Mesozoic era, the Barents Sea region was part of an extensive epicontinental sea on the northern margin of the Pangea supercontinent, with wide and shallow branches stretching into the present Siberia, the Pechora, West Siberian and Khatanga/Anabar basins (Figs. 2.15, 2.16 and 2.17).

In this Late Jurassic epicontinental sea the average water depth was from 300 to 500 m and mainly low lying land areas surrounded the basins (Fig. 2.15). Within

the distant west and northwest, the Arctic basin widened and opened towards the Panthalassa Ocean (paleo-Pacific). In the Arctic basin shallow marine, clastic sedimentation dominated during the Jurassic, as elaborated in the stratigraphical presentation above. The low relief sea floor was disrupted by a limited number of topographic highs, where coarser grained sedimentation took place. Coarse-grained sedimentation also happened along the coastlines and in the near-shore environments of this extensive basin. Eustatic sea level changes controlled the overall major depositional conditions, while heavy winds and storms triggered periods with sand deposition in this normally rather calm, epicontinental sea. Consequently, the general sedimentation in the basin was dominated by deposition of fine grained clays, with sparse amounts of sand and silt. The low relief surrounding land areas suffered rather heavy chemical weathering and dominating well-weathered dissolved and fine-grained material was transported into the basin. The Barents Sea region was at that time located in a position of about 50° N-latitude and clearly experience more chemical weathering and both warmer and more humid conditions than of today. Both along the coastal margins of the mainland of Norway and the Greenland region of that time, shallow marine and highly reworked, well weathered quartzitic sands were deposited. In periods and regions with reduced clastic sedimentation, carbonates and organic matter dominated.

Due to the wide and shallow bathymetry of this extensive Late Jurassic-earliest Cretaceous sea, the total organic production must have been substantial. In the water-masses large amounts of planktonic algae thrived and provide good living conditions for a variety of animals higher up in the food-chain. This is partly reflected in periods with high accumulation of fossil skeletal fragments from ammonites, belemnites, mollusks and marine reptiles such as ichtyosaurus and plesiosaurs, which are commonly found preserved along shallow banks and highs in this sea. The *Dorsoplanites* Beds of Svalbard is a good example of such skeletal accumulations (Fig. 2.19) (Dypvik et al. 1991a).

The sedimentation and preservation of organic matter was prolific, as shown on Svalbard by the black shales of the Janusfjellet Subgroup, in the western Barents Shelf by the Hekkingen Formation and in the Timan-Pechora region by the organic rich Bazhenov Formation. In the epicontinental sea, high algal production took place, while rich vegetation must have been present in the surrounding land areas. Large parts of this organic matter were eventually deposited on the sea floor. Due to lack of global glaciations in the Late Jurassic, oceanic circulation was reduced and only sporadic storms existed. The limited ocean-circulation resulted in rather sluggish to partly stagnant bottom water conditions on the Barents Shelf. Consequently, deposited organic matter was rarely altered or oxidized, and came to make up a large part of the bottom sediments. Some of these dark, organic rich clays (e.g., parts of the Hekkingen Formation) have been found to contain more than 20 wt% total organic carbon (Dypvik et al. 1991a; Leith et al. 1992).

The wide paleo-Arctic seaways in the Jurassic provided a good basis for establishing inter-basinal stratigraphical correlations, since comparable beds were deposited over wide areas and the deposition in the Circum-Arctic basins was mainly controlled by sea level changes (Figs. 1.3, 2.5 and 2.16) (Mørk and Smelror

Fig. 2.19 Ammonites in the Dorsoplanites Bed of the Agardhfjellet Formation at Knerten, near Wimanfjellet, on the south shore of Isfjorden, Svalbard. The ammonite is *Dorsoplanites maximus* of Middle Volgian (Photo: Hans Arne Nakrem)

2001). Consequently the Circum-Arctic geological successions, we study today contain several stratigraphical markers that can be traced across wide areas. In the Mjølnir case, this makes the search, discovery and dating of ejecta layers possible.

Figure 2.15 shows the Kimmeridgian and Volgian palaeogeographic architecture of the Arctic. The Oxfordian setting was overall transgressive. This transgressive development and widening of the epicontinental sea is evident in the Canadian Arctic (Balkwill 1978; Embry 1991) and in the Siberian basins. The evolution may be related to the Late Jurassic opening of the North Atlantic and the initial sea floor spreading. The Valanginian reconstruction (Fig. 2.16) displays regressive developments into the Early Cretaceous of the Canadian and central Arctic basin, while the deepening of the North Atlantic takes place in combination with a tectonic uplift of the northern paleo-Barents Sea.

The Late Jurassic to Early Cretaceous setting with widespread fine-grained sedimentation lasted for 15–20 million years. The succeeding Middle and Upper Cretaceous formations generally is characterized with several additional stratigraphic breaks and non-continous sequences, and more coarse grained sedimentation filling up the subsiding basins.

Early Cretaceous (Barremian-Aptian) volcanic activity was widespread on Kong Karls Land and Franz Josef Land and the offshore surroundings areas. The volcanics are represented by subalkaline tholeiites (Amundsen et al. 1998). In the Svalbard and the northern Barents Sea dolerites of Late Jurassic to Early Cretaceous ages are present (Halvorsen 1989; Halvorsen et al. 1996).

The paleogeographic setting during the Volgian-Ryazanian boundary times when the Mjølnir bolide hit the paleo-Barents Sea is shown in Fig. 2.15. The bolide approached the target site at an angle of 45° from the south/southeast (Tsikalas 2005). No doubt that the 1.5–2 km in diameter bolide created a major disturbance

in this Jurassic/Cretaceous world. The effects of the impact, its related mass flows and great tsunamis are described in detail in the following chapters of this book.

During the 20–25 million years time-span from the beginning of Oxfordian to end of the Valanginian several impactors should have hit (based on periodicity and statistical estimates) the extensive paleo-Greenland –Barents Sea – Eastern Siberian epicontinental seas. Numerous tsunamis should have formed, and their traces preserved in the sedimentary record, e.g., as coarse grained and exotic deposits with sharp erosional features along the palaeo-coasts lines. In the deeper, more central parts of the basin, it may be difficult to see tsunami effects, in contrast to the shallow banks or platforms of the Arctic Sea. These deeper regions could consequently be well suited for crater search, both due to the low and rather continuous accumulation rates of fine-grained clastics and the well developed fossil floras and faunas making detailed stratigraphical correlation possible.