Why is Svalbard an island? Evidence for two-stage uplift, magmatic underplating, and mantle thermal anomalies

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[1] Svalbard is an anomalous, subaerial part of the Barents Shelf, Northeast Atlantic Ocean. In this study, we performed both, one- and two-dimensional subsidence analyses based on basin structure, water depth, and thermochronology, to quantify and date the phases of uplift affecting Svalbard during the Cenozoic. Svalbard has experienced two phases of uplift, from >36 to ~10 Ma, and since ~10 Ma, similar in timing to uplift phases identified in Greenland, Scandinavia, and the Barents Shelf. Total uplift across much of the Central Tertiary Basin of Svalbard is >1.5 km and exceeds 2.5 km in parts of the West Spitsbergen Foldbelt (WSFB). Uplift from >36 to ~10 Ma accounts for the greatest part of the vertical motion and like the younger phase reduces in magnitude towards the east. Flexural rigidity of the lithosphere is estimated to be low (\(T_e \approx 5 \text{ km}\)), so that post-36 Ma erosion of the WSFB contributes little to the uplift, whose permanent nature and proximity to the synchronous Yermak Plateau favors a link to regional magmatic underplating. Plume dynamic support and flexural unloading along the western transform plate margin can be ruled out as influences on vertical motions. Since ~10 Ma renewed uplift, generating the modern topography may be linked to thermal erosion of the mantle lithosphere under Svalbard. We suggest that a likely cause of much of the surface uplift is the northward propagation of the Knipovich Ridge to establish continuous seafloor spreading through the Fram Strait after ~10 Ma.


1. Introduction

[2] The vertical motions of passive continental margins are controlled by the nature of strain accommodation in the crust and mantle lithosphere, which generally results in long-term subsidence in settings of extension [McKenzie, 1978; Karner and Watts, 1982]. Nonetheless, it has also been observed that in many circumstances passive margin subsidence can be delayed or even reversed into kilometer (km)-scale uplift [Japsen et al., 2006; Green and Duddy, 2010; Japsen et al., 2010; Japsen et al., 2012], or is anomalous in some other respect compared to simple models. There have been a number of explanations advanced to explain these anomalies. Rock uplift along passive margins has been noted during active extension, and this has variously been linked to flexural uploading of the rifting margin (i.e., an elastic response of the plate to the stresses of extension), and to thermal erosion (i.e., thinning of the lithosphere as a result of increased temperatures in the asthenosphere) or preferential extension of the mantle lithosphere [Royden and Keen, 1980; Mutter et al., 1988; Steckler et al., 1988].

[3] Many non-volcanic rifted margins show more subsidence than would be expected for the observed degrees of upper crustal extension, implying a depth-dependent extension (i.e., different parts of the plate extend to different degrees) [Davis and Kuszni, 2004]. In this context, upper crust is taken to mean that part that deforms in a brittle fashion, lying above the ductile-to-brittle transition, typically 10–15 km deep [Wernicke, 1981; Zuber et al., 1986]. Volcanic rifted margins, in contrast, have long been recognized to exhibit subsidence anomalies, which have been ascribed to mantle thermal anomalies that are linked to the increased melting. This may produce two types of uplift. Crustal thickening by magmatic addition to the base or within the crust causes uplift that does not decay with time. Additional temporary uplift may be related to upwelling asthenosphere due to its lower than normal density, and this in turn can result from either thermal anomalies or compositional differences, e.g., caused by melt extraction [White and McKenzie, 1989; Phipps Morgan et al., 1995].

[4] Permanent subsidence anomalies linked to magmatic underplating have been noted around the North Atlantic basin and are generally interpreted to reflect excess melting during breakup, accentuated by the presence of a thermal anomaly, probably the Iceland mantle plume [Brodie and
2. Geological Setting

[7] Svalbard forms the northwestern edge of the Barents Shelf and is one of few emergent areas along its northern margin. The archipelago is positioned immediately adjacent to the actively spreading ocean basins of the Fram Strait and the Eurasian Basin (Figure 1). The Arctic Ocean is subdivided by the Lomonosov Ridge into the late Mesozoic-early Cenozoic driven by a series of tectonic events [Blythe and Kleinspehn, 1998; Dörr et al., 2012]. In particular, these include Cretaceous passive margin uplift of the Amerasian Basin and the Paleogene Eurekan deformation as a result of the formation of the Eurasian Basin and the Norwegian-Greenland Sea [Blythe and Kleinspehn, 1998; Dörr et al., 2012]. The appearance of Miocene basalts unconformably overlying erosion surfaces in northern Svalbard requires that Svalbard was subaerial no later than the Miocene. Erosion surfaces in central Svalbard may be of the same age or could reflect uplift following late Cenozoic glaciation of Svalbard and the Barents Shelf. Nevertheless, the question remains which processes caused permanent uplift of Svalbard and what provides the basis for the modern topographic appearance of the archipelago.

[6] In order to quantify the amounts and timing of uplift, and to isolate the controlling mechanisms, we model the vertical motions of Svalbard as recorded by the sedimentary rocks of the CTB and the present topography. We particularly focus on the Cenozoic history in this study, building on earlier work on the WSFB [Blythe and Kleinspehn, 1998].

2.1. Post Breakup History

[8] The younger tectonic history of the northwestern Barents Shelf goes back to the breakup of Pangaea, marked by the formation of the Atlantic Ocean (~175 Ma) [Klitgord and Schouten, 1986], which initiated the redistribution of modern Arctic landmasses [Lawver et al., 2002]. The formation of the Amerasian Basin and associated magmatism dominated the tectonic history of Svalbard in Jurassic-Cretaceous time [Vogt et al., 1982; Lane, 1997] and led to exhumation of the northern Svalbard provinces [Dörr et al., 2012].

[9] Late Cretaceous-Paleocene spreading in Baffin Bay-Labrador Sea [Roest and Srivastava, 1989; Chalmers and Pulvertaft, 2001], overlapping with and followed by spreading in the Norwegian-Greenland Sea, moved Greenland north. As a consequence of this, deformation of northern Greenland and Ellesmere Island formed the Paleogene Eurekan Orogen [De Paor et al., 1989], and the WSFB in Svalbard [Tessensohn and Piepjohn, 2000; Saalmann et al., 2005]. At the same time, Svalbard is situated immediately adjacent to two active rift zones in a location relatively remote from the effects of the Iceland plume (Figure 1). Thus, it offers a chance to look at the vertical motion of a passive margin of the slow-spreading Arctic Ocean and Norwegian-Greenland Sea. The very presence of the archipelago raises the question as to why it is there at all, because all the surrounding extended crust of the Barents Shelf lies below sea level. Svalbard, by contrast, is subaerial and locally mountainous (up to 1700 m high), with the topography of the Central Tertiary Basin (CTB) almost surmounting the West Spitsbergen Foldbelt (WSFB) (Figure 2). Elevated ancient erosion surfaces are also noted in central and northern Svalbard, specifically in Andrée Land and Ny-Friesland, where they are observed to tilt gently towards the north [Harland, 1969; 1997]. Earlier studies have demonstrated diachronous exhumation of Svalbard in late Mesozoic-early Cenozoic driven by a series of tectonic events [Blythe and Kleinspehn, 1998; Dörr et al., 2012].
deformation of western Svalbard triggered subsidence in the adjacent CTB where km-scale Paleocene-Eocene strata are preserved [Dallmann, 1999]. The WSFB then experienced exhumation of about 2–3 km since the Eocene-Oligocene [Blythe and Kleinspehn, 1998].

Seafloor spreading in the Eurasian Basin to the north of the Barents Shelf commenced ~55 Ma [Kristoffersen, 1990], and this process separated the Lomonosov Ridge from the northern Barents Shelf [Brozena et al., 2003]. Enhanced magmatism during the Eocene formed a large igneous province (LIP) within the Arctic Ocean, although this was separated by Oligocene breakup into the Yermak Plateau (to the northwest of Svalbard) and the Morris Jessup Rise [Srivastava and Roest, 1989; Müller and Spielhagen, 1990] (Figure 1).

Figure 2. Map of Svalbard showing the geological units and major structural features discussed in this study. Map is modified from Dallmann et al. [2002]. Inset map shows the four major areas of Svalbard mentioned in the text. Red dashed lines show depths of burial of the Paleocene Firkanten Formation as estimated by Manum and Throndsen [1978].

[10] Seafloor spreading in the Eurasian Basin to the north of the Barents Shelf commenced ~55 Ma [Kristoffersen, 1990], and this process separated the Lomonosov Ridge from the northern Barents Shelf [Brozena et al., 2003]. Enhanced magmatism during the Eocene formed a large igneous province (LIP) within the Arctic Ocean, although this was separated by Oligocene breakup into the Yermak Plateau (to the northwest of Svalbard) and the Morris Jessup Rise [Srivastava and Roest, 1989; Müller and Spielhagen, 1990] (Figure 1).

[11] Pronounced northward-propagating seafloor spreading in the North Atlantic and in the Eurasian Basin finally led to the separation of Svalbard from Greenland. Seafloor spreading commenced in the Fram Strait in Oligocene time (anomaly 13) [Tessensohn and Piepjohn, 2000; Engen et al., 2008]. The youngest preserved strata on Svalbard are basalts in NW Spitsbergen and Ny-Friesland dated at 10.4 ± 1.1, 11.5 ± 1.2, and 8.7 Ma by K-Ar methods [Prestvik, 1978; Tebenkov and Sirotkin, 1990; Dallmann, 1999]. Minor amounts of Quaternary basalts have also been reported in NW Svalbard [Amundsen et al., 1988; Skjelkvåle et al., 1989] (Figure 2). The Miocene basalts extruded on an erosion surface of basement units and remnants of Devonian sandstones [Gjelsvik and Ilyes, 1991]. The Barents Shelf underwent exhumation since the Late Eocene, and most of
2.2. The Topographic and Morphologic Appearance of Svalbard

Svalbard can be subdivided into four main areas (northern and eastern Svalbard, the WSFB, and the CTB), which are marked by significant topographic and morphologic differences. In northern Svalbard, where mainly basement units and Devonian sedimentary rocks are exposed, relief increases inland from a low-elevation morphology in the north and northeast to higher amplitudes in central and northwestern Svalbard. The highest elevations of Svalbard (up to 1713 m) can be found in southern Ny-Friesland, not within the WSFB, as might be expected given that this is the area that has most recently suffered tectonic deformation (Figure 2). Despite the pronounced morphology, remnants of an elevated pre-Miocene erosion surface and overlying Late Miocene basaltic units are preserved in northwestern Svalbard and in Ny-Friesland [Prestvik, 1978; Harland, 1997; Dallmann, 1999]. Erosion surfaces and basalts indicate gentle pre-Late Miocene relief in northern Svalbard, although incision and pronounced relief formation is also known to post-date 10 Ma because the basalts of that age are eroded and uplifted. The surface tilts gently to the north, and in Andrée Land, there is an elevation difference of 300 m across 30 km of erosion surface.

Elevated erosion surfaces are also expressed by the flat table-top mountains of the CTB, which represents the inverted foreland basin of the WSFB. These surfaces have a typical elevation of ~800 m above sea level. Long-term erosion of the CTB and additional late Cenozoic incision formed the modern morphology. In contrast, western Svalbard, where mainly basement units are exposed, is more mountainous, but lower in average elevation above sea level compared to the CTB.

The Triassic to Paleogene units of the CTB were generally deposited in shallow marine shelf or coastal and deltaic facies. The succession is coarsening upward from mainly shales and siltstones to mainly fine-grained sandstones from the Sassendalen, Kapp Toscana, Adventdalen to Van Mijenfjorden Group. The Adventdalen Group is further subdivided into the Janusfjellet Subgroup and the Helvetiafjellet and Carolinefjellet Formations, the latter separated from the Paleogene Van Mijenfjorden Group by a major hiatus.

3. Analytical Methods

3.1. One-Dimensional Backstripping

The vertical motions of the CTB can be quantified by subsidence reconstructions, the most basic style of which involves analysis of a hypothetical borehole in the middle of the CTB derived from the geological map and cross section of Dallmann et al. [2002]. The sedimentary section was “backstripped” using the standard methods of Sclater and Christie [1980], which assume local isostatic compensation. This type of analysis calculates the depth to basement at any given time in the past after correcting for the loading effects of the sedimentary cover, its burial compaction, and the water depth of deposition. The objective is to isolate a purely tectonically driven subsidence record. The greatest uncertainties in this analysis are in the water depth estimates, although these are minor in this case because the sedimentary rocks are all shallow water or terrestrial deposits. A further correction was made for long-term, first-order sea-level variations using the reconstruction of Haq et al. [1987]. We employ the sediment compaction parameters of Sclater and Christie [1980] taken from North Sea drilling data to correct for the dewatering of the sediments on burial. This should be a reasonable approximation in the absence of any information that indicates an abnormal burial and compaction history for the CTB. The estimated thicknesses and average lithologies are shown in Table 1. Because there is no actual borehole in the center of the CTB, our analysis is based on a hypothetical section, often referred to as a “pseudo-well” in that location which is designed to understand the vertical tectonics in a place where the subsidence is at a maximum. For this reason, many thicknesses used in the analysis reflect the maximum values on the island. Similarly, because there are significant facies changes across the basin our estimates of the lithological composition of each group represent a simplification. Errors introduced into the analysis by this simplification will result in changes in the magnitude but not the timing of the vertical motions.

3.2. Two-Dimensional Subsidence Modeling

Because of the assumed local isostatic compensation, this approach is necessarily in error when the lithosphere has significant flexural strength. As described below, we consider the effective elastic thickness (T_e) to be low in this region, so that the one-dimensional backstrip may be a reasonable estimate of tectonically driven basement subsidence. The most useful feature of the analysis is its ability to highlight times of accelerated basement subsidence and thus of either tectonic extension, or flexural downwarping.
using a two-dimensional method. This is superior in being about to account for lateral changes in stratal thicknesses and also to include a correction for the flexural strength of the crust under Svalbard. Although Karner and Watts [1982] proposed that the flexural rigidity of continental margins is low during active extension and then increases with time after that ceases, the strength of continental crust during the rift-drift transition has remained a matter of controversy. Subsidence evidence from many sedimentary basins suggests that flexural strength during extension is typically low [Barton and Wood, 1984; Fowler and McKenzie, 1989; Watts and Stewart, 1998]. While some workers have argued for a weak continental lithosphere, with strength located principally in a brittle, relatively thin, upper crust layer [Maggi et al., 2000], others have invoked significant flexural strengths in rift zones [Ebinger et al., 1989; van der Beek, 1997], based on gravity studies of the flexural wavelength around the rift, as well as the presence of seismic activity deep in the plate [Jackson and Blenkinsop, 1993; Foster and Jackson, 1998]. A general division can be made between the rifting of thermally mature, cratonic continental lithosphere and hot, young, orogenic, or subduction-altered crust, representing strong and weak end members, respectively. From the long-term tectonism that has affected the Svalbard region, we anticipate a low flexural strength in this setting.

[18] We estimate flexural rigidity under Svalbard by considering the width of the Cenozoic foreland basin. We use the cross section of Dallmann et al. [2002] in our analysis but correct for the erosion that the CTB has experienced during the Cenozoic because only a fragment of the original basin has been preserved (Figures 2 and 3). Manum and Throndsen [1978] used vitrinite reflectance data from Nordenskiöldfjellet (on the northern rim of the CTB) and consideration of the overall southward-plunging geometry of the CTB to infer the stratal thickness and estimated that around 3.5 km of sediment were originally preserved at the basin center. This number is not especially well defined, and could be less, since only 1.9 km is preserved in the southern-central CTB. However, our result is not strongly affected by the effective thickness because an overestimate would result in an overstated degree of observed subsidence anomalies, while these anomalies would not be eliminated if the eroded amount is less than proposed by Manum and Throndsen [1978]. Figure 3 shows our interpretation of the stratigraphy and proposed eroded section derived from the thickness of the eroded section and consideration of the preserved structure in eastern Svalbard. Based on that reconstruction, the modern outcrop width of the Van Mijenfjorden Group of ~30 km is suggested to have been ~70 km in the original at the end of the West Spitsbergen Orogeny. A simple flexural equation allows us to assign a Te of 4.9 km to the Svalbard crust at the time of flexure [Turcotte and Schubert, 1982]. In order to determine the robustness of our results, we run models with a Te of 2 km (basin width of 40 km) and of 10 km (basin width of 135 km), as well as the preferred 5 km value in this study.

[19] In order to determine subsidence anomalies we restore the CTB to its state in the past and compare this with known elevations determined from the stratigraphy. The unloading of the CTB was simulated using the FlexDecomp software of Badley Geoscience Limited, which has been applied to a variety of basin analysis studies [Kuszmir et al., 1995], including flexural foreland basin studies [Clift and VanLaningham, 2010]. This model treats the continental lithosphere as an elastic beam in a normal fashion. For each model, the eroded material was assigned the same granitic density as the uneroded mountains, i.e., 2750 kg/m³. The model was run for three possible effective elastic thicknesses (Te) of 2, 5, and 10 km.

4. Results

4.1. One-Dimensional Backstripping

[20] The one-dimensional backstripping exercise predicts the depth to basement through time after correction for the
tion since foreland Van Mijenfjorden Group. The lack of sedimentation continued and reached a maximum rate in the Late Jurassic (~157 Ma) and continued through the Cretaceous until 99 Ma after which a lengthy hiatus means that the history spanning 99–65 Ma is missing. Nonetheless, we know that subsidence continued and reached a maximum rate in the Paleogene, synchronous with the sedimentation of the foreland Van Mijenfjorden Group. The lack of sedimentation since >36 Ma [Cepek and Krutzsch, 2001] means that we have no control on the vertical motions after that time, beyond the fact that the basin has experienced large-scale inversion before the present day, as demonstrated by the modern topography and the exposure of the Paleozoic-Cenozoic sedimentary fill that requires an end to subsidence and burial.

4.2. Two-Dimensional Modeling Results

[21] In this study, we focus on the timing and extent of basin uplift during the Cenozoic history of Svalbard. To do this, we restore the CTB to its state prior to inversion and then estimate the effects of different tectonic processes in controlling the vertical motions. We have estimated the pre-inversion structure and stratigraphic thicknesses in Figure 3. However, from the fluvial and deltaic facies of the Van Mijenfjorden Group, we know that this unit was deposited effectively at sea level, hence adjacent to the deforming WSFB and not at higher elevations (Figure 5). This seems reasonable because even in the central Himalayan foreland basin the modern elevations are close to sea level.

[22] If this structure can be considered to be a reasonable reconstruction of the CTB at >36 Ma, then we can estimate the effect of eroding away the topography of the WSFB that would have been higher than the top of the foreland basin at that time. In a first model, we presume that after formation of the foldbelt, erosion would have been concentrated in the orogen and not in the flat lying basin. Unloading of the crust by the removal of this mass might be expected to generate uplift, not just of the deep-buried rocks in the foldbelt, but also more regionally because of flexure. For example, such uplift and partial inversion of a foreland basin is observed to follow erosional unloading of the Himalayan foreland in the Early Miocene [Clift and VanLaningham, 2010]. The size of the eroded mountain load is estimated from the different depths of exhumation derived from fission track studies within the WSFB and CTB. According to data from Blythe and Kleinspehn [1998], around 1.5–2.0 km of additional erosion has occurred in the foldbelt compared to missing overburden estimates of the CTB since the Eocene [Manum and Throndsen, 1978].

[23] The uplift of the CTB caused by erosion of this 1.5–2.0 km from the orogen after 36 Ma is shown in Figure 6 using the range of T_e values considered appropriate. The calculation shows that uplift at the mountain front would be ~420 m for each model but the extent of the uplift varies between the models. In the T_e = 2 km model, uplift is restricted to within 21 km from the front of the WSFB, while it rises to 32 km for the preferred T_e = 5 km model, and 46 km for T_e = 10 km. Further east, all models predict modest amounts of subsidence (~120 m). Although erosion in the WSFB results in a certain amount of uplift close to the range front of the foldbelt, it is clear

![Figure 4](image-url)  
**Figure 4.** Result of one-dimensional backstripping showing the predicted depths to basement after correction for sediment loading, water depth and sea level changes, grey shaded regions show periods of hiatus. The arrow on the left side of the figure indicates the modern corrected depth to basement. Black bars indicate uncertainties in water depths. Note the basin subsided until 36 Ma marked by the top of Van Mijenfjorden Group, followed by inversion until the present day.

![Figure 5](image-url)  
**Figure 5.** Initial situation of the CTB at the end of Van Mijenfjorden Group sedimentation (>36 Ma) with the surface of the basin around sea level. Top of the West Spitsbergen Foldbelt is defined on the basis of fission track data [Blythe and Kleinspehn, 1998].
that the magnitude and width of the uplifted zone are insufficient to explain the common ~800 m modern altitude of the CTB. This result is true regardless of which $T_e$ value is employed.

[24] In a second step, we consider the effect on topography of the erosion of the WSFB plus 2.5 km of erosion from the CTB. In this stage of the model, we removed all the material that would have overlain the rocks now found in the plateau region and peaks of central Svalbard. The valleys that now incise the ~800-m-high erosion surface (which is known to be older than ~10 Ma based on the overlying volcanic rocks in northern Svalbard), however, were not included. In our analysis, we assume that the 800 m erosion surface in the CTB is the same as the one dated at >10 Ma in southern Ny-Friesland and Andrée Land which even lies at elevations of ~1000 m there. Given the lack of age data that require them to be separate surfaces, for the time being, we assume that they have a common origin. Thus, we consider the effect of long-term erosion between the end of sedimentation of the Van Mijenfjorden Group and the final formation of the ~10 Ma erosion surface. The starting condition for this stage of the model is shown in the upper panel of Figure 7, where the uplift of the top of the Van Mijenfjorden Group that was associated with initial erosion of the WSFB can be seen to decrease away from the range front. The lower panel of Figure 7 shows the predicted topography after removal of 2.5 km of CTB (which had overlain the 10 Ma erosion surface) and WSFB cover based on our preferred 5 km $T_e$ value. What is apparent is that except for a brief interval between 18 and 26 km distance on the model section, the entire width of the section lies under sea level. Because we know that this surface was formed subaerially between >36 Ma and ~10 Ma, the actual depth of the top of the section must have been close to sea level. Therefore, the predicted submarine depths represent a clear depth anomaly, and some additional process beyond flexural unloading must have caused permanent uplift that is in excess of 1 km across much of the CTB and decreases moving towards the east.

[25] In the final stage of our analysis, we assess the degree of uplift experienced by Svalbard since 10 Ma. Greenland glaciation dates from 7 Ma [Larsen et al., 1994] and the Northern Hemispheric Glaciation started in force at ~2.7 Ma [Raymo, 1994] so that much of the uplift and erosion after 10 Ma may have been accomplished under glacial conditions. In order to do this and separate any post-10 Ma depth anomalies from the earlier >36 Ma to ~10 Ma, we again correct the modeled section to sea level based on the observation that the 10 Ma erosional surface was formed close to sea level [Fjellanger and Sørbel, 2005]. In this final phase of unloading, we remove the material that would have filled the present-day valleys, i.e., the mass between the 10 Ma surface and the modern topography. As before, we run the model three times, with our chosen range of $T_e$ values (Figure 8). As might be expected with a topography starting at sea level, our models all predict that the peaks of the mountains were uplifted above sea level related to the unloading effect of the glacial erosion in the adjacent valleys. Despite this uplift component, the overall effect of removing material results in all models in a mostly subaerial topography, which is in stark contrast to the high modern topography (Figure 8). The difference between the model and the modern topography can thus be interpreted as the depth anomaly that has developed since 10 Ma. The lower panel of Figure 8 shows that this anomaly is highest within ~1 km in the region of the WSFB and reduces to the east, averaging ~450 m across the CTB.
5. Discussion

[26] The subsidence modeling described above shows that erosion of the WSFB and the CTB does not provide sufficient uplift to account for the present topographic appearance of Svalbard. Compression is not considered to be a reason for uplift for several reasons because central and western Svalbard experienced uplift after the Paleocene-Eocene formation of the WSFB and the CTB. Central Svalbard underwent exhumation after the compression in western Svalbard had ceased. Furthermore, there is no structural evidence for significant compression in Svalbard after the Late Eocene. Therefore, we consider now the timing, magnitude, and extent of the depth anomalies defined above in order to better understand how they were caused and what this tells us about the geodynamics of Svalbard. The pre-10 Ma, post-36 Ma subsidence anomaly is recognized as the largest part of the total anomaly (Figure 9). This part of the uplift shows a generally regular variability, steadily decreasing eastwards from a maximum at the 40 km mark on the modeled section. This roughly correlates with the eastern edge of the preserved CTB. West of this point, the anomaly is much more variable and decreases to less than zero under the center of the modern CTB. Surprisingly, there is even a negative anomaly close to the 20 km mark on the section close to the center of the CTB. The anomaly is then high again, but quite variable over short horizontal distances close to the CTB-WSFB contact. Some of this short wavelength variability may be caused by the deformation and folding of the western side of the CTB, and which the subsidence-backstripping analysis used here is not well equipped to correct for.

[27] The timing of the post-36 Ma uplift is broadly consistent with that reconstructed from western Greenland by Japsen et al. [2010] and may also have been related to the lateral resistance to plate motion as proposed by Japsen et al. [2012]. However, we do not favor this particular explanation to account for the bulk of these vertical motions. Although there is uplift of the shelf between Norway and Svalbard, as well as further towards the east along the Barents Sea shelf [Cavanagh et al., 2006; Green and Duddy, 2010], the amount of this uplift is clearly greater in Svalbard as evidenced by the existence of an island here and not elsewhere along the shelf edge. The thermochronologic evidence provided by Green and Duddy [2010] indicates km-scale uplift and exhumation during the latest Eocene across the Barents Shelf. However, the fact that this area is now covered by seawater demonstrates that the process...
which caused the uplift had a temporary effect, which is in contrast to our inference from subsidence analysis of Svalbard. The site of best age and vertical control used by Green and Duddy [2010] lies approximately 700 km south of Svalbard and is quite remote from the influence of the WSFB or the Yermak Plateau unlike the central basin of Svalbard. This implies that if Svalbard was influenced by this regional effect, then it was also experiencing additional processes that drove the permanent uplift. The increase in the subsidence anomalies approaching the continent-ocean transition does suggest a link between that transition and the uplift, rather than this being linked to something that influences the entire Barents Sea. The uplift in Svalbard appears to be concentrated at the corner of the continental margin. Currently, the age control is insufficient to know for sure that the uplift we reconstruct is actually of the same timing as that seen in Western Greenland. Although we remain open to the possibility of a more regional uplift that also affects Svalbard, in this study, we also look to more regional and testable triggers for tectonic uplift.

5.1. Influence of Earlier Rifting Events

One influence that we can totally discount as driving uplift is the effect of thermal subsidence following the last clear rifting phase, which we defined as having been in the Late Jurassic-Cretaceous (Figure 4). Because this extension started at 157 Ma and likely was not long-lasting, we know that the thermal subsidence that followed the rifting would be largely finished by the present. Around 500 m of basement subsidence occurred in ~60 m.y. of time (157–99 Ma) and this implies quite modest amounts of extension in the first place, i.e., ~9% or a beta factor of 1.09 by simple application of the instantaneous, uniform, pure shear extension model [McKenzie, 1978]. In any case thermal subsidence might have been expected to increase the depth to basement and so slightly increase the depth anomalies, especially for the older preglacial anomaly. However, a beta factor of 1.09 would predict only ~30 m of thermal subsidence during the period of 65 to 40 Ma, assuming a short-lived rifting event starting at 157 Ma. Because the depth anomalies are mostly ~1 km, we can conclude that residual effects from

![Figure 8](image-url)
the last rifting event to have affected the Svalbard lithosphere are not significant. In any case, thermal subsidence would cause more subsidence while our unloading exercise described above shows that the CTB and the underlying Mesozoic might be expected to still be deeply buried under the Barents Shelf rather than uplifted as a modern island.

5.2. Dynamic Mantle Uplift

Several processes are capable of driving surface uplift, but the most likely in this setting are erosion of the dense mantle lithosphere root under the archipelago, crustal thickening by magmatic underplating, and dynamic support of the lithosphere driven by the presence of underlying upwelling mantle. Because the depth anomaly, which formed between >36 Ma and ~10 Ma, is still present, it seems unlikely that dynamic support can be invoked as the cause of this uplift. Although it is theoretically possible that prior to 10 Ma, there may have been an intervening phase of reburial and subsequent renewed exhumation, there is no evidence for this having occurred. Hence, applying Occam’s razor, we model the simpler history in which the lack of net subsidence is the result of stability during this period. Dynamic support can be excluded because for this process to be effective, Svalbard would have to remain stationary over the buoyant circulating mantle asthenosphere. While Eurasia has been relatively stationary for much of the Cenozoic, there is certainly no suggestion from the present geology that the archipelago is currently located on top of a region of major mantle upwelling. Although volcanism is known from the Middle Miocene and even the Quaternary [Prestvik, 1978; Tebenkov and Sirokin, 1990; Dallmann, 1999], the volumes of these lavas are very low and not consistent with the high degrees of melting normally associated with mantle plume settings. The alkaline composition of the Quaternary volcanic rocks indicates low degrees of partial melt and is not a typical plume-type magma [Sushchevskaya et al., 2009].

The lithosphere under Svalbard is presumably not as thin as that seen in modern Iceland. That limits the degree of asthenospheric upwelling and melting under Svalbard. Apart from that, basaltic eruptions onto the rifted continental crust of the Greenland margin during Eocene breakup involved much thicker sequences, whose geochemistry point to higher degrees of partial melting [Fitton et al., 1998] compared to those of Svalbard. Furthermore, subsidence analysis of the Greenland and NW European margins has shown that dynamic support linked to plume activity was generally short lived (5–15 m.y.). This is because of the plate drift, and even plume migration, within the hotspot reference frame [Clift et al., 1998; Champion et al., 2008]. We therefore conclude that Svalbard is neither now located above a mantle plume, and anomalous elevation is not driven by dynamic support.

5.3. Role of Magmatic Underplating

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mantle, and this appears to be the sort of uplift Svalbard has experienced. The degree of uplift is dependent not just on the amount of melt added to the crust, but also its composition. Seismic velocity analysis suggests that the underplated material is typically gabbroic in composition because of its high velocity (>7.0 km/s) in such locations as East Greenland or eastern North America [Holbrook and Kelemen, 1993; Hopper et al., 2003].

The pre-10 Ma depth anomaly could be explained by magmatic underplating of Svalbard, related to the formation of the Yermak Plateau adjacent to Svalbard. Although some of the uplift may also be due to the regional plate motion-related uplift proposed by Green and Duddy [2010], the direct juxtaposition of the plateau to Svalbard and the knowledge that magmatic underplating was the cause of much continental margin uplift around the North Atlantic in the latest Paleocene-Eocene [Brodie and White, 1994; Clift and Turner, 1998] must also be considered. As noted above we do not exclude other influences in the mode of Green and Duddy [2010] but uplift is clearly greater in Svalbard than the adjoining parts of the Barents Shelf, as demonstrated by the fact that Svalbard is an island and the rest of the shelf is not. Simple isostatic considerations require that underplating must cause uplift. If the underplating was gabbroic, with a density of 3.0 g/cm³ (using a normal 3.33 g/cm³ density for the mantle), then our pre-10 Ma depth anomalies would require ~11–14 km of underplating in the eastern CTB, decreasing further east, assuming local isostatic compensation. If all of the anomaly were caused by underplating then maximum values would peak at >18 km under the WSFB, but fall to zero close to the present center of the CTB. Values of 18 km are very high, particularly in comparison to those observed under the Voring Plateau or East Greenland margins, where underplating is in the range of ~10 km close to the continental-ocean transition in locations that were not directly affected by the center of the Iceland plume [Mjelde et al., 2002]. The 11–14 km of underplating in the eastern CTB is very similar to the seismically estimated underplated range of 10–15 km observed under the Faeroe Islands [Richard et al., 1999]. Given the proximity of Svalbard to the Yermak mantle thermal anomaly, it is possible that our underplated estimate is broadly correct, although, it is noteworthy that seismic refraction surveys have not yet identified a thick zone of high velocity crust under NW Svalbard [Ritzmann and Jokat, 2003]. Why however would the anomaly fall to low values under the central CTB? The major north-south trending faults across the Svalbard archipelago may influence the distribution of underplating in ways that we do not fully account for here. However, assuming that the low anomalies described are real and not just a function of poor modeling linked to folding, one possibility is that underplating would be highest where the lithosphere was thinnest, allowing more upwelling and melting [McKenzie and Bickle, 1988]. That means that underplating would be highest on the eastern side of the preserved basin, or in reverse, that the Moho under the CTB and base of the lithosphere would be deflected downwards, potentially reducing the degree of melting and underplating. It is noteworthy that Early Cretaceous dolerite sills are thicker in the east of Svalbard than in the west [Neibert et al., 2011], which would be consistent with our proposed underplating pattern.

Underplated gabbro would be expected to have an impact on the thermal evolution of the CTB. Addition of gabbroic liquid to the crust during Paleogene magmatism would increase basal heat flow until the excess heat of the intrusion relative to the country rock has been conducted or convected away. We use the conductive cooling model of Pedersen [1993] to estimate the increase in heat flow. This model accounts for latent heat of cooling and is used to model the effect of igneous underplating and the presence of a hotter than normal asthenosphere under Svalbard. The Pedersen [1993] model has been modified so that the melt thickness in each area is not determined by assuming adiabatic melting of hotter than normal mantle during extension, because there is no extension occurring at the time of underplating. Using a background mantle-derived heat flow value of 57 mW/m², we calculate that 14 km of underplated gabbro would raise heat flow to 92 mW/m² around 12 m.y. after its initial emplacement. This represents a maximum increase in the geothermal gradient from 25°C/km to 42°C/km (using an average thermal conductivity of 2.2 W/m.K) [Midttomme and Roaldset, 1999], provided that all the magmatic addition is at the base of the crust. If magmatic bodies were emplaced at shallower crustal levels, heat flow would increase further, but might also dissipate rapidly if the intrusions were thin. Across much of the CTB underplated thicknesses would have been lower than 14 km, and at a minimum in the basin center meaning that heat flow would have been only moderately increased following magmatism.

### 5.4. Effects of Transform Tectonics

Given the lack of corroborating seismic evidence for major underplating and the spatial links between the depth anomalies and the transform plate boundary along the Hornsund Fault Complex, we investigate the possible role of strike-slip tectonics in causing uplift between >36 Ma and ~10 Ma. The timing of the depth anomaly is consistent with a link to the Yermak Plateau emplacement, but also with the formation of the transform plate boundary along the Hornsund Fault Complex (Figure 1). The decrease in depth anomaly towards the east could reflect increasing distance from the thermal anomaly and the associated reduction in melting, but could also indicate plate margin processes.

Uplift could also be related to thermal erosion of the dense mantle root under the archipelago. Thereby, active mantle asthenosphere convection is driven by the sharp horizontal thermal gradient across the Hornsund Fault Complex between the thicker, older lithosphere of Svalbard and the thinner, younger lithosphere of the Norwegian-Greenland Sea (Figure 1). Such boundaries have been proposed to drive enhanced circulation and erosion of the thicker lithosphere in passive margin settings where the continent-ocean boundary is sharp [Mutter et al., 1988]. However, this type of boundary is especially a feature of transform margins, such as west of Svalbard, where convective removal of the mantle lithosphere has been modeled to drive km-scale uplift [Todd and Keen, 1989]. Further, permanent uplift has been linked to migration of spreading centers along transform continental margins as result of the magmatic underplating that this may cause [Lorenzo et al., 1991]. The >36 Ma to ~10 Ma uplift of Svalbard, however, does not seem to have been caused by transform margin processes, because Svalbard was not juxtaposed against a spreading axis at that time. Although modeling suggests
high degrees of uplift (>2 km) where spreading axes meet transform margins, this falls away rapidly from the plate boundary (<100 km) and also decays rapidly through time [Todd and Keen, 1989]. We note that the CTB now lies >100 km from the continent-ocean boundary, so that the degree of uplift linked to spreading ridge migration would have been very low, even if the timing was appropriate. We do however recognize the fact that the continent-ocean boundary would have been much closer, ~50 km [Eldholm et al., 1987], between magnetic anomalies 13 and 24 (Oligocene). Then thermal rejuvenation as a result of the proximity to a spreading center could have been important, if one had been present. However, magnetic anomalies show that all spreading segments between the Norwegian-Greenland Sea and the Eurasia Basin were active after 10 Ma [Engen et al., 2008]. Thus, the Knipovich Ridge propagated towards western Svalbard and potentially could have driven some of the uplift in more recent times. We conclude that if transform margin processes have caused any uplift in Svalbard, then it must have been during the younger <10 Ma phase.

5.5. Erosion of the Mantle Lithosphere

[36] If magmatic underplating linked to formation of the Yermak Plateau causes much of the depth anomalies generated between >36 Ma and 10 Ma, we must also ask about those that have formed since 10 Ma (Figure 8). It is these later anomalies that explain the fact that Svalbard is a landmass and not just a submerged part of the Barents Shelf, because we know that the island was close to sea level at 10 Ma. In this respect Svalbard may be a more dramatic example of the regional uplift that is seen in East Greenland, the North Slope of Alaska, and across the Barents Sea at the same time [Green and Duddy, 2010]. A second phase of magmatic underplating seems less likely to account for this uplift, because the Quaternary volcanic rocks of northern Svalbard are very limited in extent and thickness. Additionally, there is no corresponding tectonic event such as the emplacement of a LIP along the margin of Svalbard at that time.

[37] Some authors, however, discussed the possibility of a rejuvenation of the “Yermak Plume” [Feden et al., 1979], or Yermak Plateau [Crane et al., 1988], resulting in elevated heat flow on the plateau. Okay and Crane [1993] further suggested that this elevated heat flow was linked to a rejuvenation of the Spitsbergen Shear Zone, caused by propagation of asymmetric spreading from the Knipovich Ridge to the south of Svalbard. In this model, the elevated heat flow would be driven by mantle upwelling associated with a broad simple shear extensional system crossing both the Yermak Plateau and Svalbard. The decrease in post-10 Ma uplift moving east across Svalbard is consistent with the depth anomalies being linked to processes active along the Spitsbergen Shear Zone. Extension by itself, however, tends to cause subsidence. So if this process is driving the uplift of Svalbard then it would have to be a preferential extension removing more mantle lithosphere than crust. Because mantle lithosphere is denser than either the asthenosphere or the continental crust, it acts as a dense root to the plate. If this dense root is preferentially removed, then the remaining buoyant crust would be uplifted. However, uniform pure shear thins both the lithosphere and the crust the same degree and results in subsidence. As a result uplift can only be achieved if strain accommodation was depth dependent and preferentially affected the mantle lithosphere. The simple shear system favored by Okay and Crane [1993] for the Yermak Plateau has potential to preferentially remove mantle lithosphere from under Svalbard, but in this case the system would have to cross the continent-ocean boundary to the north of Svalbard.

[38] Thinning of the mantle lithosphere under Svalbard would provide a ready mechanism to explain this most recent phase of uplift [Vågnes and Amundsen, 1993]. The location of Svalbard at the corner of the continental lithosphere of the Barents Shelf allows such mantle circulation to potentially operate on both the north and west side of the archipelago. Exactly why lithospheric erosion commenced after 10 Ma is not clear, but, it could be linked to the establishment of the continuous seafloor spreading axes through the Fram Strait as the Knipovich Ridge extended further to the NW [Engen et al., 2008]. Alternatively, the lithospheric erosion might have increased as a result of the resistance to regional plate motion that may have caused synchronous uplift in the Barents Sea, East and West Greenland, as well as the North Slope of Alaska [Green and Duddy, 2010]. Perhaps, the location of Svalbard at the edge of the continental lithosphere caused a stronger reaction compared to the adjoining parts of the Barents Sea. We can at least say that there is no linkage to deep-seated mantle plume activity or to rifting tectonics, as is the case with some other North Atlantic uplifted terranes [Balkwill, 1987; Rohrman et al., 1995].

6. Conclusions

[39] In this study, we present a series of subsidence reconstructions across central Svalbard and show that the observed depths of sedimentation are often not consistent with depths predicted from our understanding of the tectonic processes. In particular, we know that prior to 10 Ma erosion of the WSFB and CTB would have resulted in a basin that should have been submarine, but was in reality close to the sea level. This requires that the archipelago had been uplifted between the end of foreland basin sedimentation and 10 Ma. The causes for this permanent uplift are unclear, but seem most likely linked to magmatic underplating during the formation of the Yermak Plateau. The lack of spreading ridge migration and the distance of the CTB from the continent margin preclude transform margin processes as being the trigger of uplift. Likewise, erosion of the WSFB causes only a very moderate flexural uplift that cannot explain the inversion of the size and extent identified. The low flexural rigidity of the lithosphere under Svalbard (T_e 5 km) precludes flexural unloading following separation of Svalbard and Greenland being a factor. Likewise, plume-type dynamic support or thermal erosion of the mantle lithosphere cannot be invoked to account for the depth anomalies generated between >36 Ma and 10 Ma because the uplift is permanent and does not dissipate through time.

[40] The second phase of uplift post-dating 10 Ma formed the modern topography of Svalbard by uplift and incision of the previously formed erosion surface. That is coeval with relief development in other parts of the Arctic, such as Greenland and Scandinavia. Our study showed that at least in Svalbard, this uplift episode cannot be linked to underplating.
or plume-type dynamic support. This phase was smaller than the earlier, but still reached ~1 km across much of the WSFB and ~450 m in the CTB. Like the earlier depth anomalies, these decrease to the east. We concur with earlier proposals that this reflects thermal erosion of the mantle lithosphere under Svalbard [Vågnes and Amundsen, 1993]. We rule out plume activity as being the cause of the thermal erosion and favor a linkage to the establishment of a continuous seafloor spreading system through the Fram Strait at that time. Thermal conditions and mantle circulation linked to the new seafloor spreading center may be responsible for thinning the mantle lithosphere under Svalbard and allowing uplift to occur. This explains why Svalbard remains as a high standing archipelago and is under Svalbard and allowing uplift to occur. This explains system through the Fram Strait at that time. Thermal conditions activity as being the cause of the thermal erosion and favor a

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