INTRODUCTION

Clinoforms represent accretionary strata that are characterized by topset, foreset and bottomset geometries, marking the transition from shallow to deeper waters. They are commonly used to interpret the changes in sedimentary environments and relative sea level in frontier basins, since they represent ‘frozen’ palaeobathymetric profiles (Patruno, Hampson, Jackson, & Whipp, 2015; Patruno & Helland-Hansen, 2018; Pirmez, Pratson, & Steckler, 1998; Sangree & Widmier, 1978; Steel et al., 2008; Steel, Olsen, Armentrout, & Rosen, 2002). Depending on their scale and sedimentary environments, clinoforms can be classified into the following types (Figure 1a; Helland-Hansen & Hampson, 2009; Patruno & Helland-Hansen, 2018): (a) Shoreline or deltaic clinoforms, which are generally produced by the progradation of deltas, barrier-islands, shorelines and strandplains. These clinoforms are normally up to a few tens of meters in height (low relief) and are typically formed over periods of 0.1–1 Myr; (b) Shelf-edge clinoforms, which display heights of hundreds of meters (high relief) and are accreted in periods of 0.1–20 Myr as result of shelf-margin progradation; and (c) Continental...
marginal clinoforms, forming sets thousands of meters high, which represent the transition from continent to ocean, and are accreted in periods of 5–100 Myr.

During the last several decades, both academia and industry have carried out detailed studies of clinoform successions to decipher the evolution and infill of sedimentary basins, since this analysis is crucial to the understanding of the distribution of reservoirs, seals and source rocks from the shoreline to the basin floor (Dreyer, Whitaker, Dexter, Flesch, & Larsen, 2005; Helland-Hansen & Hampson, 2009; Houseknecht, Bird, & Schenk, 2009; Rønnevik, Beskow, & Jacobsen, 1982; Ulmishek, 2003). Different techniques have been adopted to analyse the distribution of coarse-grained sediments within clinoforms:

1. Sequence stratigraphy, which is based on the analysis of stacking patterns, geometric relationships and stratal terminations, to identify key surfaces formed as result of relative sea level fluctuations (Catuneanu et al., 2011; Mitchum, Vail, & Sangree, 1977). These key surfaces representing breaks in sedimentation divide the sedimentary succession into genetic units with chronostratigraphic significance.

2. Trajectory analysis, which is based on the study of vertical and lateral migration of clinoforms and associated sedimentary environments resulting from the interplay between sediment input, bathymetry, eustatic sea level fluctuations and tectonics (Bullimore, Henriksen, Liestøl, & Helland-Hansen, 2005; Helland-Hansen & Gjelberg, 1994; Helland-Hansen & Hampson, 2009). This method has been commonly used for outcrops and seismic reflection profiles to understand the changes in palaeoenvironmental conditions and associated lithological distribution within shoreline clinoforms (Hernández-Molina et al., 2000; Marin, Escalona, Nøhr-Hansen, Śliwińska Kasia, & Mordasova, 2017; Patruno et al., 2015), shelf-edge clinoforms (Glørstad-Clark, Faleide, Lundschiøen, & Nysetuen, 2010; Johannessen & Steel, 2005; Marin et al., 2017; Poyatos-Moré et al., 2016; Steel et al., 2002) and continental margin clinoforms (Salazar, Moscardelli, & Wood, 2016, 2018). Trajectory analysis of shelf-edge clinoforms is based on the description of the rollover point migration (i.e. shelf-edge), which can be classified into ascending, flat or descending trajectory (Helland-Hansen & Hampson, 2009). Flat and descending shelf-edge trajectories are formed as result of a stable or falling relative sea level, respectively, and commonly display oblique progradational seismic patterns (Figure 1c,d). These trajectories indicate less storage potential on the shelf due to marine and subaerial erosion, and subsequent sediment bypass to the slope and basin floor (Carvajal & Steel, 2006; Johannessen & Steel, 2005). On the other hand, ascending shelf-edge trajectories are formed under a long-term rise in relative sea level, and generally show sigmoidal progradational seismic patterns (Figure 1b). Due to the continuous rise in relative sea level, these trajectories favour higher sediment storage on the shelf than in the slope and basin floor (Helland-Hansen & Hampson, 2009). However, in the case of narrow shelves with anomalously high sediment supply, or lack of wave-remobilization of sediment at the shoreline, shoreline or deltaic clinoforms can reach the shelf-edge forming shelf-edge deltas, delivering significant amount of sediment to the slope and basin floor even during the periods of relative sea level rise (Carvajal, Steel, & Petter, 2009; Carvajal & Steel, 2006; Dixon, Steel, & Olariu, 2012; Jones, Hodgson, & Flint, 2015).

3. Geometric analysis of clinoforms involves studies such as the measurement of foreset angles (Mitchum et al., 1977; Patruno et al., 2015; Pirmez et al., 1998). A relationship between foreset angles and lithology has been suggested based on the observations of modern deltas, outcrops and analogue models (Anderson, Chidsey, Ryer, Adams, & McClure, 2004; Nemec, 2009; Orton & Reading, 1993; Patruno et al., 2015; Pirmez et al., 1998). These studies conclude that sandy clinoforms tend to have steeper foreset angles than fine-grained clinoforms (Figure 1e,f; Adams & Schlager, 2000; Anderson et al., 2004; Orton & Reading, 1993; Pirmez et al., 1998). However, other factors can also affect the foreset angles, including: (a) basin physiography and tectonics, where the foreset angles tend to increase during progradation into deeper waters (Figure 1h; Klausen et al., 2018; Orton & Reading, 1993; Pirmez et al., 1998; Porąbski & Steel, 2003; Ross, Halliwell, May, Watts, & Syvitski, 1994; Steckler, Mountain, Miller, & Christie-Blick, 1999); (b) sediment input; where the foreset angles tend to increase in areas with low sediment

---

**Highlights**

- Salt mobilization causes drastic vertical and lateral changes in relative sea level, which in turn induce lateral variations in clinoform trajectory, foreset angle, relief and progradation rates.
- Salt withdrawal and uplift produces complex spatial and temporal stacking patterns of depositional environments resulting in different palaeogeographies through time.
- The use of forward stratigraphic models is essential in tectonically active areas, since the isolated use of conventional methodologies in clinoform analysis might conduct to interpretation pitfalls such as wrong interpretation of trajectories and overestimation of foreset angles.
- This study has implications in understanding sediment partitioning of salt-bearing basins filled by prograding overburdens.
input and decrease in areas with high sediment input (Coleman & Wright, 1975; O’Grady, Syvitski, Pratson, & Sarg, 2000); (c) relative sea level, where periods of relative sea level rising lead to steepening foreset angles (Figure 1g; Pirmez et al., 1998; Ross et al., 1994); and (d) turbidity currents, which tend to decrease the foreset angles (Kostic, Parker, & Marr, 2002; Pratson et al., 2007).

Previous studies have mostly analysed the geometry and trajectory of clinoforms in passive margins (Anderson, 2005; Anderson et al., 2016; Steckler et al., 1999), foreland basins (Pellegrini et al., 2017; Steel et al., 2002), back-arc basins (Salazar, Moscardelli, & Wood, 2016, 2018) and epicontinental seas (Eide, Klausen, Katkov, Suslova, & Helland-Hansen, 2017; Glørstad-Clark et al., 2010; Klausen et al., 2018; Klausen, Ryseth, Helland-Hansen, Gawthorpe, & Laursen, 2015; Marin et al., 2017; Riis, Lundschienn, Høy, Mørk, & Mørk, 2008). However, the literature also provides few examples of shelf-edge clinoforms prograding through basins with ongoing halokinesis such as salt-related passive margins (e.g. Gulf of Mexico and Santos Basin; (Ge, Jackson, & Vendeville, 1997; Jackson, Jackson, & Hudec, 2015; Koyi, 1996), salt-related foreland basins (e.g. Paradox Basin; Trudgill, 2011) and salt-related rift basins (e.g. Nordkapp Basin; Koyi, Talbot, & Tørudbakken, 1995; Rojo & Escalona, 2018; Rowan & Lindsø, 2017). Contrary to other
basins without the presence of salt, salt-bearing basins show a complex feedback between the salt and the sediments. On one hand, prograding overburdens cause differential loading on the underlying salt and trigger halokinesis (Ge et al., 1997; Koyi, 1996; Trudgill, 2011). On the other hand, salt movement generates spatial and temporal variations in palaeobathymetry. These variations consisting of uplift (e.g. salt diapirs) and subsidence (e.g. minibasins) influence the progradation of clinoforms and the spatial and temporal distribution of sedimentary environments (Banham & Mountney, 2013; Rojo & Escalona, 2018). Among these two salt–sediment interaction processes, the effect of prograding overburdens is well-understood by both analogue and numerical models (Albertz & Ings, 2012; Ge et al., 1997; Koyi, 1996; Warszatka, Kley, & Kukowski, 2013) and has been studied in many salt-bearing basins (Ge et al., 1997; Jackson et al., 2015; Koyi, 1996; Trudgill, 2011), whereas few studies discuss the influence of salt movement on prograding clinoforms (Cohen & Hardy, 1996) or other similar geometries such as sand dunes (Kopriva & Kim, 2015; Piliouras, Kim, Kocurek, Mohrig, & Kopp, 2014).

The study area includes part of the Norwegian Barents Shelf, which displays a complex distribution of rift basins (some of which are salt-bearing), structural highs and platforms (Figure 2a; Faleide, Vågnes, & Gudlaugsson, 1993; Worsley, 2008). Along its geological history, the Barents Shelf has experienced two main episodes of clinoform progradation; (a) during the Triassic (Figure 2b) and (b) during the Cretaceous (Figure 2c), where clinoforms prograded across the entire shelf. Most previous studies have analysed Triassic (Eide et al., 2017; Glørstad-Clark et al., 2010; Klausen et al., 2018, 2015; Riis et al., 2008) and Cretaceous clinoforms (Grundvåg et al., 2017; Marin et al., 2017) in tectonically stable areas (e.g. Bjarmeland and Finnmark platforms). However, few studies (e.g. Heiberg,

FIGURE 2  (a) Main structural elements of the Barents Sea. The Tromsø, Nordkapp and Tydlibanken basins are salt-related basins with abundant diapiric structures. Inset map shows the location of the Barents Sea in the Arctic region (after Rojo et al., 2019). (b) Palaeogeography during the Late Triassic in the Barents Shelf. Shelf-edge clinoforms prograded towards the NW of the Barents Sea and across salt-related basins (e.g. Tiddlybanken and Nordkapp basins) with ongoing halokinesis (modified from Henriksen, Bjørnseth, et al., 2011). (c) Palaeogeography during the Early Cretaceous (modified after LoCrA (2018)). Shelf-edge clinoforms prograded towards the southern part of the Barents Sea and across salt-related basins with ongoing halokinesis. Black arrows in maps B and C indicate the progradation direction
2018) have been carried out in salt-related basins (e.g. Nordkapp or Tiddlybanken Basins) where salt tectonics may have influenced clinoform progradation and sedimentary environments. Therefore, this study combines seismic interpretation, 3D structural restoration, and forward stratigraphic modelling of the Tiddlybanken Basin to accomplish the following objectives: (a) decipher the impact of salt movement on the geometry and trajectory of Lower Cretaceous clinoforms and (b) understand the implications of clinoform geometry and trajectory variations on sediment partitioning in the basin. The study area provides an excellent opportunity to understand how salt movement triggers the formation of complex basin physiography. This results in local changes in clinoforms geometry and trajectory, whose analysis is crucial for understanding the infill and evolution of the basin, as well as predicting potential reservoirs.

2 | GEOLOGICAL SETTING

The Tiddlybanken Basin is a 50 km-wide, NW–SE elongated salt basin between the Finnmark platform and the Fedynsky High (Figure 2a). A 30-km long, NW–SE trending salt wall is present at the basin axis, whereas an E–W salt-cored anticline, known as the Signalhorn dome, defines the southwestern basin boundary (Figure 3a; Mattingsdal, Høy, Simonstad, & Brekke, 2015). The Tiddlybanken Basin is considered a frontier basin without exploration wells, since for decades it was part of a politically disputed area between Norway and Russia until the border agreement was established in 2011 (Gernigon et al., 2018; Rowan & Lindsø, 2017). The geological history of this basin has been strongly influenced by tectonic events and climate variations that affected the entire Barents Sea (Gernigon et al., 2018, 2014; Henriksen, Ryseth, et al., 2011; Worsley, 2008).

2.1 | Tectonostratigraphy

2.1.1 | Late palaeozoic

The Tiddlybanken Basin formed during the Late Devonian–Mississippian, as the result of extension and reactivation of the major Timanian-Caledonian structures (NW–SE magnetic trends) and basement-related structures of the Norwegian Barents Shelf (Gernigon et al., 2018, 2014; Rowan & Lindsø, 2017). Based on outcrop observations in Spitsbergen (Worsley, 2008) and shallow wells in the Finnmark platform (Bugge et al., 1995), it is interpreted that pre-salt strata may consist of Mississippian-Pennsylvanian, synrift alluvial to fluvial deposits interbedded with coal (Billefjorden Gp.; Figure 3a,b).

During the Pennsylvania–early Permian, extensional faulting ceased and the basin underwent a period of post-rift subsidence, evolving as a sag basin until the end of the Palaeozoic (Gernigon et al., 2018; Rowan & Lindsø, 2017). The northward movement of Pangea towards arid latitudes allowed the deposition of warm-water carbonates and platform evaporites at basin shoulders and platforms (e.g. Finnmark platform), whereas thick, late syn rift-early post-rift halite-rich evaporite sequences precipitated in the basin axis (Gipsdalen Gp; Figure 3a, b; Gernigon et al., 2018; Henriksen, Ryseth, et al., 2011; Rowan & Lindsø, 2017; Stemmerik, Elvebakk, & Worsley, 1999; Stemmerik & Worsley, 2005). During the late Permian, continuous northward movement of Pangea towards colder latitudes accompanied by ongoing passive subsidence favoured the deposition of cool and cold water carbonates with spiculites (Figure 3a, b; Henriksen, Ryseth, et al., 2011; Stemmerik et al., 1999; Stemmerik & Worsley, 2005). During the Early Middle Triassic, contraction associated with the development of the Uralides to the east triggered salt mobilization in the Tiddlybanken Basin (Rowan & Lindsø, 2017). Resulting accommodation from salt withdrawal was then filled by NE–SW striking transgressive-regressive fluviodeltaic sediments sourced from the Uralides and Fennoscandia (Sanssendalen Gp.; Figures 2b, and 3a,b; Glørstad-Clark et al., 2010; Klausen et al., 2015, 2018; Lundschiien, Høy, & Mørk, 2014). During the Late Triassic-Jurassic, minibasin subsidence and diapir uplift decreased dramatically causing the burial of the NW–SE salt wall at the basin axis (Figure 3a,b; Rowan & Lindsø, 2017). This decrease in accommodation was experienced in most of the Barents Sea and resulted in the deposition of condensed, shallow marine-fluvio deltaic deposits with complex drainage systems (Kapp Toscana Gp.; Figure 3a,b; Anell, Brøthen, & Olaussen, 2014; Henriksen, Ryseth, et al., 2011).

The Late Jurassic–Early Cretaceous period was marked by passive subsidence with minor faulting possibly associated with the opening of the North Atlantic in the western Barents Sea (Figure 3a,b; Faleide et al., 2008, 1993). Late Jurassic regional flooding resulted in the deposition of fine siliciclastics and organic shales in most of the Barents Sea (Adventalen Gp.; Figure 3a,b; Anell, Brøthen, & Olaussen, 2014; Henriksen, Ryseth, et al., 2011). Uplift in the northern part of the Barents Sea during the Barremian (Early Cretaceous) induced a new regression of the shoreline and shelf deposits towards the south (Adventalen Gp.; Figure 2c; Grundvåg et al., 2017; Marin et al., 2017). This episode was followed by Aptian–Albian sea level rise and subsequent basin infill with shelf deposits (Figure 3a,b). Thinning of the Lower Cretaceous strata towards the salt wall in the Tiddlybanken Basin indicates that diapir growth was likely rejuvenated due to the differential loading induced by these prograding Lower Cretaceous sediments towards the south of the shelf (Figures 2c, and 3a,b; Gernigon et al., 2018; Heiberg, 2018).

2.1.2 | Cenozoic

During the Late Cretaceous–Cenozoic, the basin underwent several contractional events caused by plate reorganizations...
FIGURE 3  (a) N–S regional seismic line through the Tiddlybanken Basin showing the seismic units and sequences mapped in this study (See Figure 3d for colour indexing). (b) Lithostratigraphic chart illustrating the main stratigraphic units, depositional environments and tectonics events affecting the Barents Sea and the Tiddlybanken Basin (modified after Larsen et al. 2002). B.P = Bjarmeland platform; Nk. B = Nordkapp Basin; F.P = Finmark platform; Tb.B = Tiddlybanken Basin. Inset map shows the location of the seismic line.
associated with the opening of the North Atlantic (Faleide et al., 2008, 1993). Consequently, the salt wall was squeezed and rejuvenated by contraction (Figure 3a,b; Gernigon et al., 2018; Rowan & Lindso, 2017). Finally, successive events of Cenozoic uplift and erosion, including the Pleistocene glacial erosion, removed ca. 1.5 km of Cenozoic and Cretaceous strata in the Tiddlybanken Basin (Figure 3a,b; Baig, Faleide, Jahren, & Mondol, 2016; Henriksen, Bjørnseth, et al., 2011; Ohm, Karlsen, & Austin, 2008).

3 | Dataset

We use one exploration well and 2D full-stack reflection data provided by the Norwegian Petroleum Directorate (NPD; Figure 3, inset map). The 2D reflection seismic survey covers an area of 54,150 km², including part of the Bjarmeland platform, eastern Nordkapp Basin, Finnmark platform and most of the Tiddlybanken Basin. The 2D seismic lines have a spacing of 4.5 km and maximum two-way travel time (TWT) of 9,216 ms. The Cretaceous interval shows frequencies ranging between 30 and 60 Hz, which allows detailed analysis of Lower Cretaceous clinoforms in terms of scale, geometry and trajectory.

Since the Tiddlybanken Basin is a frontier area, the closest exploration well-available (7229/11-1) in the dataset is located in the Finnmark platform ca. 100 km west of the study area (Figure 3, inset map). Well data consist of a conventional suite of wireline logs (e.g. gamma ray, caliper, neutron, density, sonic and resistivity), check-shots and well tops, which were integrated to: (a) provide age constraints and correlate the main seismic units and sequences across the basin, (b) assign a lithology to the clinoforms and (c) construct a 3D velocity model for time-to-depth conversion of the surfaces resulting from seismic interpretation.

4 | Methodology

4.1 | Stratigraphic framework

Based on reflection amplitude, strata terminations and continuity of seismic events, eight key horizons were identified for the regional interpretation of the Tiddlybanken Basin and nearby platforms (Figure 3a,b). The ages of these horizons are determined from well data, which were tied to the seismic interpretation. More information regarding interval velocities, K factors (change of interval velocities with depth) and densities of the different intervals is given by Clark et al., 2014; Gernigon et al., 2018; Rojo et al., 2019. The 3D structural restoration of the Tiddlybanken Basin was performed to: (a) show the progressive evolution of salt structures and palaeobathymetries through time; (b) remove salt-related deformation and compaction effects before measuring the angles and trajectories of the Lower Cretaceous clinoforms; and (c) provide uplift and subsidence maps (mm/yr) for S4 sequence, which are the tectonic input for the forward stratigraphic modelling. During the restoration process, regional levels were defined on the Finnmark platform, which does not show major salt-related deformation. The eroded and missing Cretaceous and Cenozoic strata were reconstructed based on previous publications (Baig et al., 2016; Henriksen, Bjørnseth, et al., 2011; Ohm et al., 2008). The reconstructed Cenozoic folding above salt diapirs was then unfolded using 3D flexural-slip to remove the shortening during this period. The Lower Cretaceous sequence boundaries S2 to S4 were restored as well with 3D flexural-slip to remove the salt-related deformation and avoid length losses in upturned strata close to salt diapirs (Rowan, 1996; Rowan & Ratliff, 2012). Sediments were decompacted using the method of Sclater and Christie (1980) since it fits well the porosity versus depth curves observed on borehole data in the Barents Sea (Klausen & Helland-Hansen, 2018). For the isostatic compensation of loads, we use flexural isostasy since there are large thickness variations along the basin and wedge-shape geometries characterizing the Lower Cretaceous sequences. An elastic thickness of 20 km was assumed based on Gac, Klitzke, Minakov, Faleide, and Schech-Wenderoth (2016).

4.2 | Structural restoration

We use the velocity model of Rojo, Cardozo, Escalona, and Koyi (2019) to depth-convert the surfaces from the seismic interpretation. More information regarding interval velocities, K factors (change of interval velocities with depth) and densities of the different intervals is given by Clark et al., 2014; Gernigon et al., 2018; Rojo et al., 2019. The 3D structural restoration of the Tiddlybanken Basin was performed to: (a) show the progressive evolution of salt structures and palaeobathymetries through time; (b) remove salt-related deformation and compaction effects before measuring the angles and trajectories of the Lower Cretaceous clinoforms; and (c) provide uplift and subsidence maps (mm/yr) for S4 sequence, which are the tectonic input for the forward stratigraphic modelling. During the restoration process, regional levels were defined on the Finnmark platform, which does not show major salt-related deformation. The eroded and missing Cretaceous and Cenozoic strata were reconstructed based on previous publications (Baig et al., 2016; Henriksen, Bjørnseth, et al., 2011; Ohm et al., 2008). The reconstructed Cenozoic folding above salt diapirs was then unfolded using 3D flexural-slip to remove the shortening during this period. The Lower Cretaceous sequence boundaries S2 to S4 were restored as well with 3D flexural-slip to remove the salt-related deformation and avoid length losses in upturned strata close to salt diapirs (Rowan, 1996; Rowan & Ratliff, 2012). Sediments were decompacted using the method of Sclater and Christie (1980) since it fits well the porosity versus depth curves observed on borehole data in the Barents Sea (Klausen & Helland-Hansen, 2018). For the isostatic compensation of loads, we use flexural isostasy since there are large thickness variations along the basin and wedge-shape geometries characterizing the Lower Cretaceous sequences. An elastic thickness of 20 km was assumed based on Gac, Klitzke, Minakov, Faleide, and Schech-Wenderoth (2016).

4.3 | Forward stratigraphic modelling

We use forward stratigraphic modelling to analyse the morphodynamics of prograding sediments influenced by salt-related subsidence and/or uplift. Specifically, we use the Geological Process Modelling (GPM) Petrel plugin (Schlumberger), which is a sedimentary simulation and visualization package that models the erosion, transport and deposition of clastic sediments based on physical equations.

4.4 | Restoration

Structural restoration of the Tiddlybanken Basin was performed to: (a) show the progressive evolution of salt structures and palaeobathymetries through time; (b) remove salt-related deformation and compaction effects before measuring the angles and trajectories of the Lower Cretaceous clinoforms; and (c) provide uplift and subsidence maps (mm/yr) for S4 sequence, which are the tectonic input for the forward stratigraphic modelling. During the restoration process, regional levels were defined on the Finnmark platform, which does not show major salt-related deformation. The eroded and missing Cretaceous and Cenozoic strata were reconstructed based on previous publications (Baig et al., 2016; Henriksen, Bjørnseth, et al., 2011; Ohm et al., 2008). The reconstructed Cenozoic folding above salt diapirs was then unfolded using 3D flexural-slip to remove the shortening during this period. The Lower Cretaceous sequence boundaries S2 to S4 were restored as well with 3D flexural-slip to remove the salt-related deformation and avoid length losses in upturned strata close to salt diapirs (Rowan, 1996; Rowan & Ratliff, 2012). Sediments were decompacted using the method of Sclater and Christie (1980) since it fits well the porosity versus depth curves observed on borehole data in the Barents Sea (Klausen & Helland-Hansen, 2018). For the isostatic compensation of loads, we use flexural isostasy since there are large thickness variations along the basin and wedge-shape geometries characterizing the Lower Cretaceous sequences. An elastic thickness of 20 km was assumed based on Gac, Klitzke, Minakov, Faleide, and Schech-Wenderoth (2016).

4.5 | Forward stratigraphic modelling

We use forward stratigraphic modelling to analyse the morphodynamics of prograding sediments influenced by salt-related subsidence and/or uplift. Specifically, we use the Geological Process Modelling (GPM) Petrel plugin (Schlumberger), which is a sedimentary simulation and visualization package that models the erosion, transport and deposition of clastic sediments based on physical equations.
To start the sedimentation model, it is necessary to input the undeformed and backstripped surface resulting from the 3D restoration, which represents the pre depositional palaeobathymetry of the sequence of interest, one of a series of parameters associated with the time zero in the model (Table 1). Other parameters include information about the location of the source and type of sediment, as well as the areas and rates of uplift and subsidence (mm/yr). It is important to note that GPM does not reproduce the effect of differential loading on salt. To reproduce this effect, we created a time-dependent tectonic function in which values of subsidence and uplift are increased locally when clinoforms reach the salt-related basin (Table 1). This tectonic function was adjusted until the simulated clinoforms geometries resembled those on the seismic profiles. Clinoform progradation was modelled using the diffusion and steady flow equations (Table 1). The diffusion equation defines the rate at which sediments move downslope proportionally to the slope gradient (Flemings & Grotzinger, 1996). Consequently, the topography or bathymetry becomes smoother over time. The steady flow equation simulates erosion, transport and deposition of sediments (Tetzlaff and Schafmeister, 2007, their equations 5 and 6). At each single point of the simulation grid, the algorithm calculates the transport capacity from the flow depth and velocity.

Erosion occurs when the flow contains less sediment than it can transport. Conversely, deposition takes place when the sediment carried by the flow exceeds the flow’s transport capacity. Values of sediment input (mm/yr) and water velocities (m/s) were adjusted to generate shelf-edge clinoforms similar to those in the seismic profiles, with realistic accretion times (0.1–20 Myr) and progradation rates (1–100 m/kyr), as described by Patruno and Helland-Hansen (2018).

5 | THE INFLUENCE OF SALT TECTONICS ON PROGRADING CLINOFORMS: THEORETICAL CONSIDERATIONS

The purpose of this section is to illustrate the potential influence of salt tectonics on prograding clinoforms based on a synthetic GPM model coupled with tectonics (Table 1, model 1 and Figure 4a). The model presented here represents a submarine platform 100 km-wide, 170 km-long and below sea level (maximum water depth = 400 m). It consists of a stable platform to the west and a salt wall bounded by two salt minibasins to the east. Diffusion and steady flow simulate constant sediment input during a period of 10 Myr. Each reproduced shelf-edge clinoform represents a time interval of

| TABLE 1 | Input parameters used in the synthetic model (Figure 4) and the case study in the Tiddlybanken Basin (Figure 13). The models were run using the GPM Petrel plugin. Sediment source abbreviations in the last column are: crsd: coarse sandstone, fsd: fine sandstone, silt: silt, and cly: clay |

| INPUT PARAMETERS |
|------------------|------------------|------------------|------------------|
|                  | Paleobathymetry  | Subsidence and uplift map | Tectonic function | Sedimentary processes |
| MODEL 1          |                  |                  |                  | Sediment diffusion |
| (Synthetic model)|                  |                  |                  | Diffusion function  |
|                  |                  |                  |                  | \(
| MODEL 2          |                  |                  |                  | Steady flow |
| (Case study)     |                  |                  |                  | Sediment diffusion |
|                  |                  |                  |                  | Diffusion function  |

(Sediment source: 23%crrsd, 25%fsd, 25%silt, 25%cly  
Water velocity: 0.04 m/s  
Transport coefficient: 0.0001  
Sea level: constant)
0.25 Myr. An uplift area of 5,000 km² in the north acts as the main sediment source. The salt wall uplift rate is from 0.08 to 0.15 mm/yr, which is within the rising rates of buried salt diapirs (Jackson & Hudec, 2017). Salt withdrawal minibasins have a subsidence rate ranging between 0.07 and 0.13 mm/yr, whereas sea level remains constant during the simulation. It is important to notice that clinoforms are progressively rotated as they prograde through subsiding minibasins and rising diapirs (e.g. Figure 4d,e, steps III–VIII). Therefore, foreset angles and trajectories are measured before post-depositional rotation (Figure 5a,b).

Section 1 in the west is across the stable platform, which does not experience subsidence or uplift associated with salt tectonics (Figure 4b). This section shows the successive progradation of sigmoidal clinoforms with a general decrease in foreset angles as they become more fine-grained from the source (Figure 4c steps I–VIII, and Figure 5b). This part of the model is not affected by salt tectonics and the clinoforms display a flat trajectory of the rollover point during the simulation (Figure 4c, step VIII, and Figure 5a, green line). Progradation rates decrease from 20 m/kyr close to the source to 13 m/kyr in the distal parts of the model (Figure 5c).

Middle Section 2 covers the edge of a subsiding minibasin surrounded by two stable platform areas (Figure 4b). The flat clinoforms trajectory of Section 1 (Figure 4c, step VIII and Figure 5a, green line) changes laterally into an ascending trajectory as it enters Section 2 where the rollover point moves upwards in response to basin subsidence and increase in relative sea level (Figure 4d, steps II–V and Figure 5a, blue line). It is important to note that this ascending trajectory within the minibasin allows the preservation of more coarse-grained sediments in the topsets in comparison with that in Section 1 (Figure 4c,d, step VIII). Similar to Section 1, sigmoidal clinoforms in Section 2 decrease in foreset angle as they become more distal from the source. However, clinoforms in Section 2 display relatively higher foreset angles (0.2–0.5° higher) due
to minibasin subsidence and increase in water depth (Figure 5b, blue in comparison to green line). Minibasin subsidence also causes lateral variations in progradation rates, which are lower in areas with salt withdrawal due to the longer time required for the sediments to fill the increasing accommodation space (Figure 5c, blue in comparison to green line).

Section 3 in the east covers a rising salt wall flanked by two subsiding minibasins (Figure 4b). In this setting, relatively higher subsidence rates along the most proximal minibasin produce the retrogradation of the system and landward movement of the rollover point (Figure 4e, steps I–III, and Figure 5a, red line). Consequently, in this minibasin, distal and fine-grained sediments overlay proximal coarse grain sediments previously deposited on clinoform topsets (Figure 4e, steps I–III). As minibasin subsidence decreases with time, the system progresses again following an ascending trajectory (Figure 4e, steps III–V). Similar to Section 2, sigmoidal clinoforms prograding through the minibasin exhibit higher foreset angles than those in Section 1 due to increasing water depth by minibasin subsidence (Figure 5b, red line). As prograding clinoforms approach the rising diapir, water depth decreases and clinoforms reduce in foreset angle from 2.8 to 1.6° (Figure 5b, red low). Salt wall uplift causes a drop in relative sea level, forming a falling trajectory and producing the subaerial/marine erosion of bypassing clinoforms (Figure 4e, steps VI–V, and Figure 5a). Note that diapir-induced erosion does not allow the preservation of the topsets (Figure 4e, steps VI–VII). During steps VI–VIII, the clinoforms reach the second subsiding minibasin, resulting in a flat to slightly ascending trajectory which finally becomes flat as the clinoforms prograde onto the stable platform (Figure 4e, steps VII–VIII and Figure 5a). In terms of progradation rates, Section 3 shows the lowest values of progradation among the three sections where negative values in the graph represent periods of retrogradation and landward movement of the rollover point (Figure 5c, red line). These differences in progradation rates in the three sections are the reason for spatial and temporal variations of the coastline and associated sedimentary environments (Figure 4a).

6 | CASE STUDY: LOWER CRETACEOUS CLINOFORMS IN THE BARENTS SEA

The case study consists of six N–S regional 2D seismic transects across the eastern part of the Norwegian Barents Sea (Figure 6, inset). The first regional transect is located in the Finnmark platform (Figure 6a), ca. at 100 km to the west of the Tiddlybanken Basin, and it intersects the well 7229/11-1. The other five regional transects cross the Finnmark platform, the Signalhorn dome and the Tiddlybanken Basin, which consists of a NW–SE salt wall surrounded by two salt withdrawal minibasins (Figure 7). In order to study the clinoforms, we have followed previous studies by Grundvåg et al., 2017; Marin et al., 2017, and as mentioned before, subdivided the Lower Cretaceous succession (SU6) into S2–S4 third-order sequences bounded by flooding surfaces (Vail,
Mitchum, & Thompson, 1977), which display a progradational pattern towards the SSW (Figure 7).

6.1 | Sequence 2 (S2), Barremian-Aptian and Sequence 3 (S3), Aptian

6.1.1 | Description

Based on well 7229/11-1 and previous studies by Marin et al. (2017), S2 includes part of the Kolje and Kolmule Formations of Barremian to Aptian age. S3 comprises the lower part of the Kolmule Formation of Aptian age (Figures 3b and 6c). The Gamma ray (GR) log through both sequences displays an erratic response, which is consistent with claystones, also observed in drilling cuttings (Figure 6c). In the seismic sections (Figures 8–11), S2 and S3 wedge out towards the west and south and are bounded by major flooding surfaces. In the Tiddlybanken Basin, S2 and S3 thin towards the central salt wall and onlap peridiapiric wedges (Figure 11). Internally, both sequences exhibit a progradational pattern towards the SSW consisting of low relief (<100m) and oblique clinoforms that show a descending trajectory (Figures 9–12, and Table 2).

6.1.2 | Interpretation

Based on previous studies by Marin et al. (2017), low relief and oblique clinoforms in S2 and S3 are interpreted as shoreline or deltaic clinoforms since they display a relief less than 100 m (Helland-Hansen & Hampson, 2009; Figures 9–11, and Table 2). Descending trajectories in the Finnmark platform indicate that clinoforms were deposited under decreasing relative sea level during a forced regression. Peridiapiric wedges and thinning of growth strata towards the salt wall provide evidence of salt growth during the deposition of both sequences (Figure 11).

6.2 | Sequence 4 (S4), Aptian–Albian

6.2.1 | Description

Sequence 4 consists of the Kolmule Formation of Aptian–Albian age (Marin et al., 2017). This sequence was drilled by well 7229/11-1 and is delimited at its base and top by major flooding surfaces (Figure 6). S4 thickens towards the NNW of the Finnmark platform and thins towards the SW (Figure 7). It shows large thickness variations in the Tiddlybanken Basin, increasing in thickness in salt minibasins and thinning.
towards the salt wall (Figures 7–11). Internally, S4 consists of high relief clinofoms (>100 m). The GR log within the clinofoms displays an erratic response that corresponds to intervals of semi-continuous, low amplitude reflections in the seismic (Figure 6). Based on drilling cuttings, these intervals consist mostly of claystones. Spikes of high density and low GR are additionally encountered within this sequence and correlate with high amplitude foresets in the seismic (Figure 6). Drilling cuttings from these intervals indicate the presence of thin dolomite layers interbedded within thick packages of claystones (Figure 6). An important characteristic of the high relief clinofoms in S4 is their large variability in height (150 to 600 m), foreset angles (0.5–5.4°) and trajectories that change along the NW–SE axis of the Finnmark platform and Tiddlybanken Basin (Figures 7–11 and Table 2). In the Finnmark platform, high relief clinofoms display heights between 150 and 350 m, low foreset angles (0.5–1.5°) and a slightly ascending trajectory (0.24°). At the minibasin edge, high relief clinofoms show higher foreset angles (1.3–2.5°) and a flat to slightly descending trajectory (−0.06°; Figure 9 and Table 2). To the east, high relief clinofoms increase in height from 150 to 600 m and foreset angles vary from 0.7 to 5.4°, as they prograde from the Finnmark platform into the salt minibasins (Figures 10 and 11, and Table 2). High relief clinofoms (500–600 m) in the minibasins are often onlapped by a wedge of low amplitude, semi-continuous to chaotic seismic reflectors (Figures 10 and 11). The clinofoms trajectory also changes, showing a flat to descending trajectory (−0.1°) at the Finnmark platform followed by a moderately ascending trajectory (0.6–1.2°) within the salt minibasins (Figures 10 and 11). It is also important to notice that the clinofoms decrease in foreset angle as they approach the central salt wall. However, their trajectories above the salt wall are not observed since S4 is eroded and truncated above this structure (Figure 11).

### 6.2.2 Interpretation

High relief clinofoms in S4 are interpreted as shelf-edge clinofoms since they exhibit heights of hundreds of meters (Helland-Hansen & Hampson, 2009; Marin et al., 2017; Table 2). Ascending trajectories in the Finnmark platform were formed in response to increasing relative sea level (Figure 8). Local flat to descending trajectories followed by ascending trajectories in the Tiddlybanken Basin could be attributed to salt movement underneath, which may have induced local changes in relative sea level due to minibasin subsidence and salt growth (Figures 9–11). Thinning of growth strata towards the salt wall and thickening towards the minibasins support this hypothesis (Figures 10 and 11). Changes in foreset angle are more difficult to explain since the

**FIGURE 8** (a) Uninterpreted seismic line through the Finnmark platform. (b) Seismic interpretation of the Lower Cretaceous interval, including clinofoms. (c) Depth-conversion and decompaction of B. Clinofoms in S3 are low relief (<100 m) and display a descending trajectory. Clinofoms in S4 are high relief (150–350 m), have low foreset angles (0.5–1.5°) and display a flat to slightly ascending trajectory (0.24°). Red line in inset map shows the location of the profile.
variations between the Finnmark platform and the Tiddlybanken Basin can be attributed to several reasons: (a) local changes in sediment supply, (b) different grain size, (c) changes in water depth and (d) tectonic tilting caused by salt tectonics.

7 | CRETACEOUS SALT TECTONICS IN THE TIDDLYBANKEN BASIN

The interpreted seismic profiles show evidence of Lower Cretaceous deposition influenced by salt tectonics. Hence, the aim of this section was to illustrate the progressive evolution of salt structures and associated areas of uplift and subsidence during the Cretaceous based on 3D structural restoration of the study area (Figure 12).

7.1 | Barremian–Aptian (S2)

At the beginning of the Early Cretaceous, the study area consisted of a ramp-type, shallow shelf tilted towards the SSW (Grundvåg et al., 2017; Marin et al., 2017). The first arrival of prograding sediments from S2 triggered differential loading of the salt in the Tiddlybanken Basin, causing the active growth of the salt wall and the formation of wide and shallow minibasins in the northern basin boundary (Figure 12a). During this time, the central salt wall formed a steep bathymetric high. Steep and unstable diapir flanks may have triggered the initiation of mass-transport complexes or debris flows, which explain the presence of peridiapiric wedges in S2 (Figure 11).

7.2 | Aptian (S3)

Progradational loading of Aptian sediments into the underlying salt contributed to further growth of salt structures and minibasins, creating a complex basin physiography characterized by drastic lateral variations in accommodation space (Figure 12b). On the one hand, the central salt wall continued growing, forming a smoother bathymetric high surrounded by minibasins. On the other hand, salt supply from beneath also caused the growth of a salt pillow to the west (Signalhorn dome).

7.3 | Aptian–Albian (S4)

Relatively higher sedimentation rates during the Aptian–Albian resulted in large salt withdrawal in the northern and southern
minibasins, which caused salt flow towards the central salt wall and southwestern salt pillow (Signalhorn dome; Figure 12c). These observations support the deposition of S4 controlled by salt tectonics, and explain the changes in geometry and trajectory of the shelf-edge clinoforms in this sequence (Figures 8–12).

7.4 | Cenozoic

Contraction and basin inversion rejuvenated the central salt wall and the Signalhorn dome. This event was followed by uplift and erosion of the Cenozoic and Upper Cretaceous strata. Today, the salt wall continues growing, forming a smooth bathymetric high at the sea floor (Figure 12d).

8 | SYN-TECTONIC DEPOSITION OF THE S4 ALBIAN CLINOFORMS: INSIGHTS FROM FORWARD STRATIGRAPHIC MODELLING

The forward stratigraphic model involves the reconstruction of the Albian sequence (S4) with the main purpose of understanding the syn-tectonic deposition of Albian clinoforms in the eastern part of the Norwegian Barents Sea (Figure 13 and Table 1, model 2). The model covers the Tiddlybanken Basin and part of the Finnmark platform, and it is extrapolated towards the Fedinsky High and Bjarmeland platform to simulate a sediment source in the NE (Grundvåg et al., 2017; Marin et al., 2017). Based on well cuttings from well 7229/11-1 and the low foreset angles (from 0.5 to 1.5°) observed in the Finnmark platform, the model simulates mud-dominated shelf-edge clinoforms. The palaeo-seafloor and the uplift and subsidence rates are obtained from the 3D restoration (Figure 12 and Table 1, model 2). The model runs for 5 Myr (109 to 104 Ma) and each clinoform represents a time span of 0.5 Myr. Based on Grundvåg et al. (2017); Marin et al. (2017), and the observed thinning of S4 towards the SW, we assume a sediment progradation direction towards the SW. This progradation is modelled using the diffusion and steady flow equations with parameter values as described in Table 1, model 2. Progradation rates, slope angles and heights of shelf-edge clinoforms are measured in each step along the three sections (Figure 13c–e). They are all within realistic values for shelf-edge clinoforms, as discussed by Patruno and Helland-Hansen (2018).

8.1 | Finnmark platform

Similar to the seismic profiles (Figures 8–11), the model displays SW progradation of shelf-edge clinoforms in the northern
Finnmark platform, where clinoforms gradually increase in height ranging from 140 to 310 m, foreset angle from 0.92 to 1.6° and decrease progradation rates from 28 to 5 m/kyr (Figure 13c–e, steps I–II). These gradual changes are attributed to increasing subsidence rate from 0.01 to 0.03 mm/yr in the Finnmark platform and subsequent increment in palaeowater depth towards the SW (Table 1, model 2). An interesting feature in both the interpreted seismic (Figure 14a and d) and the model (Figure 14b and e) is the resulting flat to slightly falling clinoforms trajectory. Based on the model results (Figure 13c and d), this falling trajectory is due to the rotation of the clinoforms by the progressive subsidence and tilting of the Finnmark platform during deposition of S4. Removing this rotation results in a flat to slightly ascending clinoforms trajectory (Figure 14c and f), which is consistent with the increase in accommodation space caused by the progressive tilting of the Finnmark platform.

8.2    Tiddlybanken Basin

The model illustrates that the clinoforms prograding through the Tiddlybanken minibasins experience changes in height, foreset angle and trajectory (Figure 13c–d) with respect to the Finnmark platform (Figure 13e). This is consistent with the seismic sections (Figures 8–11). Uplift and subsidence maps indicate that the subsidence rates in the Tiddlybanken minibasins range from 0.05 to 0.10 mm/yr, whereas the active growth and uplift of the diapir is approximately 0.1 mm/yr (Table 1, model 2). Higher subsidence rates by salt withdrawal, dramatically increase the water depth in the minibasins and generate steeper slopes from 1.6 to 2.8° (Figure 13c–d, steps II–V). The clinoform height is also affected by this process, increasing from 310 to 570 m in the minibasins. Increases in relative sea level associated with minibasin subsidence decreases the progradation rates from 19 to 5 m/kyr and results in moderately ascending trajectories at the edge of the minibasins (Figure 13c–d, VII), while during the same time areas in the Finnmark platform display flat trajectories and higher progradation rates (Figure 1, II–V). These lateral differences in clinoform progradation rates and trajectories among the three sections cause spatial and temporal variations of the coastline and shelf-edge (Figure 13a). Finally, as minibasins are filled with clinoform

FIGURE 11 (a) Uninterpreted seismic line through the Finnmark platform and the Tiddlybanken Basin. (b) Seismic interpretation of the Lower Cretaceous sequences and clinoforms. (c) Depth conversion, decompaction and unfolding of B. Clinoforms in S3 are low relief (<100m) and display a descending trajectory. Clinoforms in S4 are high relief (150-350m) and show higher foreset angles as they approach the minibasin (0.7–5.4°). Note the change in clinoforms trajectory from slightly ascending (0.11°) at the basin boundary to moderately ascending (1.15°) in the minibasins. Marine onlaps may indicate the presence of slope aprons deposited in minibasins. Red line in inset map shows the location of the profile

FIGURE 12 3D structural restoration of the study area illustrating salt tectonics and associated changes in palaeobathymetry from the Cretaceous to the present. Note how prograding overburdens of S2, S3 and S4 cause the mobilization of the underlying salt which creates lateral variations in accommodation space. Red polygon in inset map shows the restored area
(a) Barremian-Aptian (S2)

Prograding sediment of S2 caused the initial loading of minibasins.

Active growth of the salt wall

(b) Aptian (S3)

Growth of salt pillow (Signalhorn dome)

Progradation of S3 caused the loading of minibasins.

Active growth of the salt wall

(c) Aptian-Albian (S4)

Growth of salt pillow (Signalhorn dome)

Active growth of the salt wall

Current growth of the salt wall

Strata eroded due to Cenozoic uplift and erosion

Legend

| Code | Formation                |
|------|--------------------------|
| S4   | Aptian-Albian            |
| S3   | Aptian                   |
| S2   | Barremian-Aptian        |
| SU6  | Upper Triassic-Upper Jurassic |
| SU5  | Middle-Upper Triassic   |
| SU4  | Lower-Middle Triassic   |
| SU3  | Lower Triassic           |
| SU2  | Upper Permian            |
| SU1  | Pennsylvanian-Lower Permian |

Bathymetry (m)

0
-1000
-800
-600
-400
-200

(d) Present day
bottomsets, water depth and slope angle decreases (from 2.82 to 1.11°), and clinoforms bypass the salt minibasin and the diapir (Figure 13c–d, V–VII).

A comparison of the interpreted seismic and the model results shows that foreset angles in the Tiddlybanken minibasins measured in the seismic (2.6 to 5.4°, Figure 14a and d) are nearly similar to those in the model (1.70 to 4.27°, Figure 14b and e). However, modelling results illustrate that foreset angles are progressively rotated by minibasin subsidence (Figure 13c–d, steps I–VII), suggesting that the angles measured at the end of the simulation do not represent the depositional foreset angles. By removing the rotation of each clinoform in the model, foreset angles in the Tiddlybanken minibasins decrease from (1.70–4.27°) to (1.59 to 2.82°) (Figure 14b–c and 14e–f). The likely overestimation of foreset angles in seismic data is further addressed in the discussion section.

9 | DISCUSSION

9.1 | Impact of salt mobilization on prograding clinoforms and sediment partitioning

Clinoforms are often encountered in salt-related basins, acting in many cases as the main trigger of salt mobilization (Jackson & Hudec, 2017). Some examples where prograding overburdens play an important role in salt mobilization are the Paradox Basin (e.g. Trudgill, 2011), Gulf of Mexico (e.g. Koyi, 1996), Precaspian Basin (Volozh, Talbot, & Ismail-Zadeh, 2003) and the Nordkapp Basin (Rojo et al., 2019; Rowan & Lindsø, 2017). Lack of well data, poor seismic resolution and post-depositional rotation and erosion during salt withdrawal and uplift are possibly the reasons why clinoforms within salt-bearing basins are poorly described in the literature. High seismic resolution and relatively low post-depositional deformation of the Lower Cretaceous strata in the Tiddlybanken Basin, however, allows study of the influence of halokinesis on prograding clinoforms. Based on a multidisciplinary approach including seismic interpretation, structural restoration and forward stratigraphic modelling, this work addresses the following effects of salt tectonics on prograding clinoforms and sediment partitioning:

1. Enhancement of slope-readjustment processes. Forward stratigraphic modelling shows that in the platform, foreset angles are constant, relatively low and are in equilibrium with sediment supply and water depth (Figures 4c, and 13e). In subsiding minibasins, progradation into increasing water depths enhances the deposition of clinoforms with relatively higher relief and foreset angle (Figures 4d–e, 4,5,13b, and 4,5,13c–d). Ross et al. (1994) suggest that tectonics (e.g. normal faulting) can considerably increase slope angles, affecting the equilibrium of the clinoform
profile and triggering slope-adjustment processes such as slope erosion, sediment bypass by gravity flows and submarine fan-apron deposition. Evidence of these processes has been additionally documented by Kertznus and Kneller (2009) in the Ebro margin (offshore, western Mediterranean) and by Salazar et al. (2016) in the Tanaraki Basin (New Zealand). Following the same arguments, clinoforms prograding into a subsiding salt minibasin might experience slope-adjustment processes favouring the deposition of submarine fan-aprons. The occurrence of onlaps onto steep clinoforms in the Tiddlybanken minibasins most likely indicate the presence of muddy slope aprons due to the overstepping of the basinal profile by salt withdrawal (Figures 10 and 11). Additionally, the presence of smaller order clinoforms embedded in the shelf-edge clinoforms might indicate the presence of shelf-edge deltas, which could have been an additional mechanism for delivering sediments to the slope and basin floor during periods of relative sea level rise (Figures 10 and 11; Dixon et al., 2012; Jones et al., 2015; Muto & Steel, 2002; Porębski & Steel, 2006).

2. Creation of complex spatial and temporal variations in accommodation space. Lateral changes in accommodation can result in different progradation rates and shelf-edge trajectories, which will result in a complex spatial and temporal distribution of depositional environments (Jones et al., 2015). Forward modelling illustrates that subsidence in salt minibasins decreases the rates of seaward progradation, whereas other areas without salt withdrawal display higher progradation rates (Figures 4c–d, 4,5,13c, and 4,5,13c–e). These differences produce lateral variations in the coastline and shelf-edge along the basin (Figures 4a and 13a), resulting in a complex palaeogeography. Salt tectonics additionally contributes to the generation of different shelf-edge trajectories, which result in a wide variability of stacking patterns and depositional environments in the basin (Figures 4 and 13). Unfortunately, stacking patterns in the Tiddlybanken Basin are difficult to observe in both seismic and models due to the mud-prone character of Albian clinoforms and the lack of nearby exploration wells. Sandy clinoforms reproduced in the synthetic model (Figure 4c–e) show that moderately ascending
FIGURE 14  (a) Unfolded and decompacted seismic section from Figure 10 through the platform and minibasin. (b) Model Section 1 parallel to the progradation direction through the platform and minibasin. (c) Depositional palaeo-bathymetric profiles of each clinoform in model Section 2 before rotation. (d) Unfolded and decompacted seismic section from Figure 11 through the platform, salt wall and minibasins. (e) Model Section 2 parallel to the progradation direction through the platform, salt wall and minibasins. (f) Depositional palaeo-bathymetric profiles of each clinoform in model Section 3 before rotation.
trajectories induced by salt withdrawal result in the vertical seaward stacking of fluviodeltaic/shelf deposits in the minibasin, which act as a temporary sediment trap of the prograding system. This agrees with analogue modelling results by Kopp and Kim (2015); Liang, Kim, and Passalacqua (2016) where tectonic tilting causes channel steering and produces stacked conformable sequences of delta lobes on the subsidence side. Nevertheless, relatively higher subsidence along the minibasin might also result in aggradational or even backstepped stacking of fluviodeltaic/shelf deposits (Figure 4e). Finally, flat to slightly ascending trajectories and poor vertical stacking of fluviodeltaic/shelf deposits characterize the surrounding platforms (Figure 4c). This lateral variability in trajectories and associated stacking patterns are consistent with the observations of Rojo and Escalona (2018) in the Nordkapp Basin, where shelf and fluviodeltaic depositional environments prograde and retrograde due to lateral variations in accommodation space during the Triassic. Diapir/pillow-induced falling trajectories were also observed in the forward models (Figures 4e, and 13d). However, its identification in salt-related basins might be an issue since clinoforms could be eroded by post-depositional diapir growth. Upper Jurassic deep-water sediments deposited during a possible diapir-induced force regression have been documented by Pena dos Reis, Pimentel, Fainstein, Reis, and Rasmussen (2017); Pimentel and Pena Dos Reis (2018) in the Lusitanian Basin (Portugal), where outcrops show spectacular incised turbiditic channels occurring at one side of a large salt wall. Based on the forward models (Figures 4e, and 13d), diapir-induced force regressions may be an important process of sediment delivery to the slope and basin floor in salt-bearing basins.

9.2 The importance of forward models in the study of clinoforms in the areas of complex tectonics

Seismic studies of ancient clinoforms are well-documented in the literature (Klausen et al., 2018; Marin et al., 2017; Patruno et al., 2015; Patruno & Helland-Hansen, 2018), and are commonly used to identify the changes in relative sea level as well as depositional environments. The conventional methodology in clinoform analysis consists on flattening the sequence to a regional datum (e.g. topset or bottomset of maximum flooding surface), with the main purpose of making the clinoform tops horizontal. Clinoforms are then backstripped or decompacted incorporating isostatic compensation. The resultant geometry is a good approximation of the clinoform palaeobathymetric profile at the time of deposition (Klausen & Helland-Hansen, 2018; Patruno et al., 2015). This methodology provides good results for clinoforms deposited in areas with low and relatively constant subsidence rates such as epicontinental seas, which have later undergone post-depositional tilting or faulting (Klausen et al., 2018, 2015; Marin et al., 2017). However, in tectonically active areas (e.g. salt-related basins) where local bathymetry/topography may be modified during sedimentation, this methodology may lead to interpretation pitfalls in clinoform geometries and trajectories. One of these misinterpretations may arise when near-horizontal topsets progressively rotate to steeper angles after their deposition. For example, topsets are nearly horizontal at the top of S4 in the interpreted seismic (Figure 14a,d) and forward model (Figure 14b,e), whereas older topsets still preserve the rotation caused by the progressive tilting of the Finnmark platform. This progressive tilting may result in the wrong interpretation of forced regressions in actively subsiding basins (Fig 14b, e). A good indicator of this pitfall is the presence of well-preserved topsets, which contradict the interpreted falling trajectories (Figure 14a–e). Another misinterpretation may arise from post-depositional rotation of clinoforms by salt withdrawal, faulting or flexure, which might lead to the overestimation of foreset angles. In this study, foreset angles increase from 0.8 in the Finnmark platform to 5.4° in the Tiddlybanken minibasins (Figure 14a,d). These high foreset angles in salt minibasins could suggest the presence of sandy-prone clinoforms. However, well data from the Finnmark platform show that the clinoforms are mud-prone. Modelling results demonstrate that these measurements do not represent the depositional foreset angles and therefore are likely overestimated (Figures 13c–d, and 13,14a–f). The model also illustrates that part of this steepening, ca. 0.8 to 2.8° (Figure 13c,d, steps I–V), is due to the increasing water depth. The remaining dip change from ca. 2.8 to 5.4° (Figure 13c,d, steps V–VII) corresponds to post-depositional rotation due to withdrawal of the underlying salt. These arguments highlight the importance of using forward stratigraphic models coupled with tectonics to study clinoforms in salt-bearing basins and other tectonically active areas. The isolated use of conventional methodologies in clinoform analysis in these complex areas might lead to potential interpretation pitfalls such as wrong interpretation of trajectories and overestimation of foreset angles, which can have negative consequences for hydrocarbon exploration models.

9.3 Model limitations and future work

Although the results clearly show the impact of salt tectonics on the geometry and trajectory of clinoforms, our simulations are based on the assumption of tectonically driven stratigraphic evolution, where uplift/subsidence rates from the structural restoration are given as input, and they do not change depending on how the sediment loading is distributed over the basin. Sediment loading is incorporated to
the restorations via decompaction and flexural deformation, but it does not dynamically reproduce the loading caused by each single clinoform. Analogue models by Piliouras et al. (2014); Koprica and Kim (2015) and numerical models by Cohen and Hardy (1996) demonstrate that factors such as sedimentation rates, initial thickness of salt and salt viscosity strongly influence the width and subsidence rate of minibasins, which in turn affect the minibasin topography and bathymetry. Therefore, future work is oriented towards analysing bidirectional salt–sediment interaction via analogue modelling or numerical modelling, to further investigate salt-controlled sedimentary architectures in salt-bearing basins.

10 | CONCLUSIONS

High-resolution Albian clinoforms together with relatively simple salt-related deformation make the Tiddlybanken Basin an excellent analogue to study the impact of salt mobilization on prograding overburdens. Based on a multidisciplinary approach including seismic interpretation, 3D structural restoration and forward stratigraphic modelling, we conclude that salt mobilization affects prograding clinoforms in the following ways:

1. Salt withdrawal/uplift causes lateral variations in the coastline and shelf-break, resulting in a complex palaeogeography.
2. Salt evacuation increases slope angles, affecting the equilibrium of the clinoform profile and triggering slope-readjustment processes such as slope erosion, sediment bypass by gravity flows and submarine fan-apron deposition.
3. Lateral and temporal variations in accommodation space caused by salt tectonics may result in different clinoform trajectories and stacking patterns along the salt-bearing basin.
4. Post-depositional rotation of clinoforms in areas of salt tectonics and other tectonically active basins might lead to interpretation pitfalls such as the wrong interpretation of trajectories and overestimation of foreset angles. These pitfalls can negatively affect reservoir prediction in prograding clinoforms.

ACKNOWLEDGEMENTS

We acknowledge the Norwegian Petroleum Directorate for giving us access to the DISKOS database and for kindly providing the NPD-BA-11 2D seismic survey in the eastern Norwegian Barents Sea. Dora Marin thanks Locra and Julocra for supporting her research. We acknowledge Sergio Courtade, Per Salomonsen and Jan Tveiten at Schlumberger for their support and technical advice with GPM. We are grateful to the reviewers William Helland-Hansen, Wonsuck Kim and Joshua F. Dixon, and the editor Ronald Steel, whose constructive comments and suggestions improved the manuscript. Thanks to Schlumberger and Midland Valley for providing academic licenses of their softwares Petrel-GPM and Move respectively.

DATA AVAILABILITY STATEMENT

Data sharing is not applicable to this article as no new data were created or analysed in this study.

ORCID

Luis Alberto Rojo https://orcid.org/0000-0002-7134-6071

REFERENCES

Adams, E. W., & Schlager, W. (2000). Basic Types of Submarine Slope Curvature. Journal of Sedimentary Research, 70, 814–828. https://doi.org/10.1306/2dc4093a-0e47-11d7-8643000102c1865d
Albertz, M., & Ings, S. J. (2012). Some consequences of mechanical stratification in basin-scale numerical models of passive-margin salt tectonics: Geological Society, London, Special Publications, 363, 303–330. https://doi.org/10.1144/sp363.14
Anderson, J. B. (2005). Diachronous development of late Quaternary shelf-margin deltas in the northwestern Gulf of Mexico: implications for sequence stratigraphy and deep-water reservoir occurrence. In L. Giosan, & J. P. Bhattacharya (eds.), River Deltas–Concepts, Models, and Examples (Vol. 83 pp. 257–276). SEPM Special Publication. https://doi.org/10.2110/pec.05.83.0257
Anderson, J. B., Wallace, D. J., Simms, A. R., Rodriguez, A. B., Weight, R. W. R., & Taha, Z. P. (2016). Recycling sediments between source and sink during a eustatic cycle: Systems of late Quaternary northwestern Gulf of Mexico Basin. Earth-Science Reviews, 153, 111–138. https://doi.org/10.1016/j.earscirev.2015.10.014
Anderson, P. B., Chidsey, J. T. C., Ryer, T. A., Adams, R. D., & McClure, K. (2004). Geologic Framework, Facies, Paleogeography, and Reservoir Analogs of the Ferron Sandstone in the Ivie Creek Area, East-Central Utah. In J. T. C. Chidsey, R. D. Adams, & T. H. Morris (Eds.), Regional to Wellbore Analog for Fluvial-Deltaic Reservoir Modeling: The Ferron Sandstone of Utah. American Association of Petroleum Geologists.
Anell, I., Braathen, A., & Olaussen, S. (2014). The Triassic-Early Jurassic of the northern Barents Shelf: A regional understanding of the Longyearbyen CO₂ reservoir. Norwegian Journal of Geology, 94, 83–98.
Baig, I., Faleide, J. I., Jahren, J., & Mondol, N. H. (2016). Cenozoic exhumation on the southwestern Barents Shelf: Estimates and uncertainties constrained from compaction and thermal maturity analyses. Marine and Petroleum Geology, 73, 105–130. https://doi.org/10.1016/j.marpetgeo.2016.02.024
Banham, S. G., & Mountney, N. P. (2013). Controls on fluvial sedimentary architecture and sediment-fill state in salt-walled minibasins: Triassic Moenkopi Formation, Salt Anticline Region, SE Utah, USA. Basin Research, 25, 709–737. https://doi.org/10.1111/bre.12022
Bugge, T., Mangerud, G., Elvebakk, G., Mork, A., Nilsson, I., Fanavoll, S., & Vigran, J. (1995). Upper Paleozoic succession on the Finnmark platform, Barents Sea. *Norwegian Journal of Geology*, 75, 3–30.

Bullimore, S., Henriksen, S., Liestøl, F. M., & Helland-Hansen, W. (2005). Clinoform stacking patterns, shelf-edge trajectories and facies associations in Tertiary coastal deltas, offshore Norway: Implications for the prediction of lithology in prograding systems. *Norwegian Journal of Geology/Norsk Geologisk Forening*, 85, 169–187.

Carvajal, C., Steel, R., & Petter, A. (2009). Sediment supply: The main driver of shelf-margin growth. *Earth-Science Reviews*, 96, 221–248. https://doi.org/10.1016/j.earscirev.2009.06.008

Carvajal, R. C., & Steel, R. J. (2006). Thick turbidite successions from supply-dominated shelves during sea-level highstand. *Geology*, 34, 665–668. https://doi.org/10.1130/G22505.1

Catuneanu, O., Galloway, W. W., Kendall, C. G. S. C., Miall, A., Catuneanu, O., Galloway, W. E., Kendall, C. G. S. C., Miall, A. (2006). Thick turbidite successions from senderal delta systems: Source-to-sink study of the southwestern Barents Sea. *GSA Today*, 16, 214–217. https://doi.org/10.1130/G22505.1

Carvajal, C., Steel, R., & Petter, A. (2009). Sediment supply: The main driver of shelf-margin growth. *Earth-Science Reviews*, 96, 221–248. https://doi.org/10.1016/j.earscirev.2009.06.008

Clift, P., Clark, J., & Hiscott, A. (2006). Sequence stratigraphy: Methodology and Nomenclature. *Newsletters on Stratigraphy*, 44, 173–245. https://doi.org/10.1127/0078-0421/2011/0011

Clark, S. A., Golorstad-Clark, E., Faleide, J. I., Schmid, D., Hartz, E. H., & Fjeldskaar, W. (2014). Southwest Barents Sea rift basin evolution: Comparing results from backstripping and time-forward modelling. *Basin Research*, 26, 550–566. https://doi.org/10.1111/bre.12039

Cohen, H. A., & Hardy, S. (1996). Numerical modelling of stratral architectures resulting from differential loading of a mobile substrate: Geological Society. *London, Special Publications*, 100, 265–273. https://doi.org/10.1144/gsl.sp.1996.100.01.17

Coleman, J. M., & Wright, L. (1975). In M. L. Broussard (Ed.), *Modern river deltas: variability and processes of sand bodies*, in (pp. 99–149). Deltas: Models for Exploration: Houston Geological Society.

Dixon, J., Steel, R., & Oliariu, C. (2012). Shelf-edge delta regime as a predictor of deep-water deposition. *Journal of Sedimentary Research*, 82, 681–687. https://doi.org/10.2110/jsr.2012.59

Dreyer, T., Whitaker, M., Dexter, J., Flesche, H., & Larsen, E. (2005). From spit system to tide-dominated delta: Integrated reservoir model of the Upper Jurassic Sognefjord Formation on the Troll West Field. *Geological Society, London, Petroleum Geology Conference, Series*, 6, 423–448. https://doi.org/10.1144/00604243

Eide, C. H., Klausen, T. G., Katkov, D., Nasuti, A. A., & Helland-Hansen, W. (2017). Linking an Early Triassic delta to antecedent topography: Source-to-sink study of the southwestern Barents Sea margin. *GSA Bulletin*, 130(1-2), 263–283. https://doi.org/10.1130/B31639.1

Faleide, J. I., Tsikalas, F., Breivik, A. J., Mjelde, R., Ritzmann, O., Eide, C. H., Klausen, T. G., Katkov, D., Suslova, A. A., & Helland-Dixon, J., Steel, R., & Olariu, C. (2012). Shelf-edge delta regime as a predictor of deep-water deposition. *Journal of Sedimentary Research*, 82, 681–687. https://doi.org/10.2110/jsr.2012.59

Ge, H., Jackson, M. P., & Vendeville, B. C. (1997). Kinematics and dynamics of salt tectonics driven by progradation. *AAPG Bulletin*, 81, 398–423.

Gernigon, L., Brönn, M., Dumas, M.-A., Gradmann, S., Grönlie, A., Nasuti, A., & Roberts, D. (2018). Basement inheritance and salt structures in the SE Barents Sea: Insights from new potential field data. *Journal of Geodynamics*, 119, 82–106. https://doi.org/10.1016/j.jog.2018.03.008

Gernigon, L., Brönn, M., Roberts, D., Olesen, O., Nasuti, A., & Yamaski, T. (2014). Crustal and basin evolution of the southwestern Barents Sea: From Caledonian orogeny to continental breakup. *Tectonics*, 33, 347–373. https://doi.org/10.1002/2013TC003439

Glorstad-Clark, E., Faleide, J. I., Lundschien, B. A., & Nystuen, J. P. (2010). Triassic seismic sequence stratigraphy and paleogeography of the western Barents Sea area. *Marine and Petroleum Geology*, 27, 1448–1475. https://doi.org/10.1016/j.marpetgeo.2010.02.008

Grundvåg, S. A., Marin, D., Kairanov, B., Śliwińska, K. K., Nohr-Hansen, H., Jelby, M. E., … Olaussen, S. (2017). The Lower Cretaceous succession of the northwestern Barents Shelf: Onshore and offshore correlations. *Marine and Petroleum Geology*, 86, 834–857. https://doi.org/10.1016/j.marpetgeo.2017.06.036

Heiberg, V. (2018). The regional Cretaceous development of the southeastern part of the Norwegian Barents Sea- from seismic interpretation. Master thesis, University of Tromsø, 116 p.

Helland-Hansen, W., & Gjelberg, J. G. (1994). Conceptual basis and variability in sequence stratigraphy: A different perspective. *Sedimentary Geology*, 92, 31–52. https://doi.org/10.1016/0037-0738(94)90053-1

Hollannd-Hansen, W., & Hamson, G. J. (2009). Trajectory analysis: Concepts and applications. *Basin Research*, 21, 454–483. https://doi.org/10.1111/j.1365-2177.2009.00425.x

Henriksen, E., Bjørnseth, H., Tals, H., Heide, T., Kyrutykhina, T., Klivjan, O., … Sollid, K. (2011). Uplift and erosion of the greater Barents Sea: impact on prospectivity and petroleum systems. In A. M. Spencer, A. F. Embry, D. L. Gautier, A. V. Stoupakova, & K. Sørrensen (Eds.), *Arctic Petroleum Geology* (Vol. 35 pp. 271–281). London, Memoirs: Geological Society. https://doi.org/10.1144/M35.17

Henriksen, E., Ryseth, A., Larsen, G., Heide, T., Romming, K., Sollid, K., & Stoupakova, A. (2011). Tectonostratigraphy of the greater Barents Sea: implications for petroleum systems. In A. M. Spencer, A. F. Embry, D. L. Gautier, A. V. Stoupakova, & K. Sørrensen (Eds.), *Arctic Petroleum Geology* (pp. 163–195). London, Memoirs: Geological Society. https://doi.org/10.1144/M35.10

Hernández-Molina, F. J., Fernández-Salas, L. M., Lobo, F., Soimoza, L., Díaz-del-Río, V., & Alveirinho Dias, J. M. (2000). The infralittoral prograding wedge: A new large-scale progradational sedimentary body in shallow marine environments. *Geo-Marine Letters*, 20, 109–117. https://doi.org/10.1007/s003670000040

Houseknecht, D. W., Bird, K. J., & Schenck, C. J. (2009). Seismic analysis of clinoform depositional sequences and shelf-margin trajectories in Lower Cretaceous (Albian) strata, Alaska North Slope. *Basin Research*, 21, 644–654. https://doi.org/10.1111/j.1365-2177.2008.00392.x

Jackson, C. A. L., Jackson, M. P. A., & Hudec, M. R. (2015). Understanding the kinematics of salt-bearing passive margins: A critical test of competing hypotheses for the origin of the Albian Gap, Santos Basin, Offshore Brazil. *GSA Bulletin*, 127, 1730–1751. https://doi.org/10.1130/B31290.1
Jackson, M. P., & Hudec, M. R. (2017). Salt Tectonics: Principles and Practice (p. 498). Cambridge University Press.

Johannesen, E. P., & Steel, R. J. (2005). Shelf-margin clinoforms and prediction of deepwater sands. *Basin Research*, 17, 521–550. https://doi.org/10.1111/j.1539-5353.2005.00278.x

Jones, G. E. D., Hodgson, D. M., & Flint, S. S. (2015). Lateral variability in clinoform trajectory, process regime, and sediment dispersal patterns beyond the shelf-edge rollover in exhumed basin margin-scale clinoforms. *Basin Research*, 27, 657–680. https://doi.org/10.1111/bre.12092

Kertzner, V., & Kneller, B. (2009). Clinoform quantification for assessing the effects of external forcing on continental margin development. *Basin Research*, 21, 738–758. https://doi.org/10.1111/j.1539-5353.2009.00411.x

Klausen, T., Aas, T. J., Haug, E. C., Behzad, A., Snorre, O., & Domenico, C. (2018). Clinoform development and topset evolution in a mud-rich delta – the Middle Triassic Kobbe Formation, Norwegian Barents Sea. *Sedimentology*, 65, 1132–1169. https://doi.org/10.1111/sed.12417

Klausen, T., & Helland-Hansen, W. (2018). Methods For Restoring and Describing Ancient Clinoform Surfaces. *Journal of Sedimentary Research*, 88(2), 241–259. https://doi.org/10.1111/jsr.2018.8

Klausen, T. G., Ryseth, A. E., Helland-Hansen, W., Gathorpe, R., & Laursen, I. (2015). Regional development and sequence stratigraphy of the Middle to Late Triassic Snadd Formation, Norwegian Barents Sea. *Marine and Petroleum Geology*, 62, 102–122. https://doi.org/10.1016/j.marpetgeo.2015.02.004

Kopp, J., & Kim, W. (2015). The effect of lateral tectonic tilting on fluviodeltaic surficial and stratal asymmetries: Experiment and theory. *Basin Research*, 27, 517–530. https://doi.org/10.1111/bre.12086

Koppriva, B. T., & Kim, W. (2015). Coevolution of minibasin subsidence and sedimentation: Experiments. *Journal of Sedimentary Research*, 85, 254–264. https://doi.org/10.1111/jsr.2015.24

Kostic, S., Parker, G., & Marr, J. G. (2002). Role of turbidity currents in setting the foreset slope of clinoforms prograding into standing fresh water. *Journal of Sedimentary Research*, 72, 353–362. https://doi.org/10.1306/081501720353

Koyi, H. (1996). Salt flow by aggrading and prograding overburdens. *Geological Society, London, Special Publications*, 100(1), 243–258. https://doi.org/10.1144/GSL.SP.1996.100.01.15

Koyi, H., Talbot, C. J., & Tørstadbakken, B. O. (1995). Salt tectonics in the northeastern Nordkapp basin, southwestern Barents sea. In M. P. A. Jackson, D. G. Roberts, & S. Snelson (Eds.), *Coarse-grained deltas* (Vol. 27, pp. 29). The International Association of Sedimentologists. https://doi.org/10.1016/0788-8589(X)900165-7

Kopp, J., & Kocurek, G. A., Mohrig, D., & Kopp, J. (2014). Triassic hydrocarbon potential in the Northern Barents Sea; integrating Svalbard and stratigraphic core data. *Norwegian Petroleum Directorate Bulletin*, 11, 3–20.

Marin, D., Escalon, A., Nøhr-Hansen, H., Śliwińska Kasia, K., & Mordasova, A. (2017). Sequence stratigraphy and lateral variability of Lower Cretaceous clinoforms in the SW Barents Sea. *AAPG Bulletin*, 101, 1487–1517. https://doi.org/10.1130/20144161001

Mattingsdal, R., Hoy, T., Simonstad, E., & Brekke, H. (2015). An updated map of structural elements in the southern Barents Sea, 31st Geological Winter Meeting, 12–14 January 2015, Stavanger.

Mitchum, R. M., Vail, P. R., & Sangree, J. B. (1977). Stratigraphic interpretation of seismic reflection patterns in depositional sequences. In C. E. Payton (Ed.), *Seismic Stratigraphy—Applications to Hydrocarbon Exploration* (pp. 117–123). AAPG Memoir.

Muto, T., & Steel, R. J. (2002). In defense of shelf-edge delta development during falling and lowstand of relative sea level. *The Journal of Geology*, 110, 421–436. https://doi.org/10.1086/340631

Nemec, W. (2009). Aspects of sediment movement on steep delta slopes. In A. Colella, & D. B. Prior (Eds.), *Coarse-grained deltas*, (Vol. 27, pp. 29). The International Association of Sedimentologists. https://doi.org/10.1016/0788-8589(00)028-e0207.CTMVOS.v2.20.CO;2

Ohm, S. E., Karlson, D. A., & Austin, T. (2008). Geochemically driven exploration models in uplifted areas: Examples from the Norwegian Barents Sea. *AAPG Bulletin*, 92, 1191–1223. https://doi.org/10.1130/0091-7613(2000)028<0207:CTMVO S>2.3.CO;2

O’Grady, D. B., Syvitski, J. P., Pratson, L. F., & Sarg, J. (2000). Categorizing the morphologic variability of siliciclastic passive continental margins. *Geology*, 28, 207–210. https://doi.org/10.1130/0091-7613(2000)028<0207:CTMVO S>2.3.CO;2

Patruno, S., & Helland-Hansen, W. (2018). Clinoforms and clinoform systems: Review and dynamic classification scheme for shorelines, subaqueous deltas, shelf edges and continental margins. *Earth-Science Reviews*, 185, 202–233. https://doi.org/10.1016/j.earscirev.2018.05.016

Pellegrini, C., Maselli, V., Gamberi, F., Asoli, A., Bohacs, K., Drexler, T. M., & Trincardi, F. (2017). How to make a 350-m-thick lowstand systems tract in 17,000 years: The Late Pleistocene Po River (Italy) lowstand wedge. *Geology*, 45, 327–330. https://doi.org/10.1130/GE03884.1

Pena dos Reis, R., Pimentel, N., Fainstein, R., Reis, M., & Rasmussen, B. (2017). Chapter 14 - Influence of Salt Diapirism on the Basin Architecture and Hydrocarbon Prospects of the Western Iberian Margin. In J. I. Soto, J. F. Flinch, & G. Tari (Eds.), *Permo-Triassic Salt Provinces of Europe, North Africa and the Atlantic Margins* (pp. 313–329). Elsevier.

Piliouras, A., Kim, W., Kocurek, G. A., Mohrig, D., & Kopp, J. (2014). Sand on salt: Controls on dune subsidence and determining salt substrate thickness. *Lithosphere*, 6(3), 195–199. https://doi.org/10.1130/L323.1
Ulmishek, G. F. (2003). Petroleum geology and resources of the West Siberian Basin, Russia. Virginia: US Department of the Interior, US Geological Survey Reston.

Vail, P. R., Mitchum, R. Jr, & Thompson, S. III (1977). Seismic Stratigraphy and Global Changes of Sea Level: Part 4. Global Cycles of Relative Changes of Sea Level: Section 2. Application of Seismic Reflection Configuration to Stratigraphic Interpretation, In C. E. Payton (Ed.), Seismic Stratigraphy - Applications to Hydrocarbon Exploration (Vol. 26, pp. 83–97). AAPG Memoir.

Volozh, Y., Talbot, C., & Ismail-Zadeh, A. (2003). Salt structures and hydrocarbons in the Pricaspian basin. AAPG Bulletin, 87, 313–334. https://doi.org/10.1306/09060200896

Warsitzka, M., Kley, J., & Kukowski, N. (2013). Salt diapirism driven by differential loading—Some insights from analogue modelling. Tectonophysics, 591, 83–97. https://doi.org/10.1016/j.tecto.2011.11.018

Worsley, D. (2008). The post-Caledonian development of Svalbard and the western Barents Sea. Polar Research, 27, 298–317. https://doi.org/10.1111/j.1751-8369.2008.00085.x

**How to cite this article:** Rojo LA, Marín D, Cardozo N, Escalona A, Koyi H. The influence of halokinesis on prograding clinoforms: Insights from the Tiddlybanken Basin, Norwegian Barents Sea. Basin Res. 2020;32:979–1004. [https://doi.org/10.1111/bre.12411](https://doi.org/10.1111/bre.12411)