Bolide Impact Origin of Norwegian Fjords

Physical Model, Erosion Mechanics, and Implications for Global Fjord Formation

Author: Ashley M. Gjøvik, B.S., J.D. (ashleymgjovik@protonmail.com) | **Date**: April 5 2026

Abstract: Norwegian fjords concentrate on Atlantic-facing coasts, achieve bedrock depths exceeding 1,500 m with no direct geochronological evidence for progressive multi-cycle glacial carving, and exhibit orientations that in the case of Sognefjord bear no clear relation to the Caledonian fold structure through which the fjord cuts (Nesje & Whillans, 1994). Glacial erosion accounts for only 35–55% of the offshore sediment volume deposited during the Plio-Quaternary in western Scandinavia (Steer et al., 2012). This paper proposes that Norwegian fjords were carved from pre-existing tectonic fracture valleys by bolide-generated mega-tsunami, with glacial modification occurring during subsequent ice occupation of the pre-existing channels.

This paper develops a physical model for the formation of the Møre og Romsdal fjords by trans-Atlantic mega-tsunami generated by the candidate Younger Dryas bolide swarm impact at ~12,800 BP (Gjøvik, 2026). The Boston Basin drumlin field — over 200 drumlins extending at least 16 km seaward into Massachusetts Bay (La Forge, 1932; Rosen & FitzGerald, 2004) — constrains the swarm to approximately 200 fragments across a ~50 km zone, with 30–50 fragments impacting shallow marine environment directly. Three independent wave generation mechanisms operate simultaneously: direct marine displacement, catastrophic meltwater release from ~8–9 km^3 of vaporized glacier, and atmospheric blast coupling at ~68× Krakatoa energy. The distributed swarm generates a quasi-planar wavefront that preserves amplitude across trans-Atlantic distance more efficiently than a point source. Energy analysis yields coastal wave heights of 15–40 m, amplifying through fjord geometry to flow velocities of 30–85 m/s — exceeding catastrophic plucking thresholds by factors of 3–17×. Dam spillway engineering data, including 90 m of bedrock scour at Kariba Dam and removal of 1,200-tonne boulders at Wivenhoe Dam at velocities below 25 m/s (Stratford et al., 2013; Bollaert, 2004), confirm that erosion at these velocity scales exceeds engineering predictions.

The pattern extends beyond Møre og Romsdal. Norwegian fjord groups have distinct orientations, each facing a water body containing a confirmed impact structure — the Barents Sea (Mjølnir, 142 Ma), the North Sea (Silverpit, 43–46 Ma), and the Atlantic (Boston Basin, proposed ~12,800 BP) — with fjord morphology correlating to impact age despite uniform gneissic bedrock across regions. The three partially submerged drumlin fields documented worldwide — Boston Harbor, Clew Bay (Ireland), and Fláajökull (Iceland) — all lie on the North Atlantic propagation path between the proposed impact site and the Norwegian coast. Globally, every major fjord coastline faces an ocean basin with documented Quaternary-age impact history. This paper presents a detailed case study of Western Norway including Møre og Romsdal and call for researchers to evaluate the mechanism at other fjord coastlines worldwide.

Citation: Gjovik, Ashley Marie. "Bolide Impact Origin of Norwegian Fjords: Physical Model, Erosion Mechanics, and Implications for Global Fjord Formation". *The Journal of Decolonized Ecology and Evolution* 1, no. 1 (April 5, 2026). https://doi.org/10.5281/zenodo.19432096.

Keywords: Norwegian fjords, Sognefjord, Hardangerfjord, Storfjord, Romsdalsfjord, Geirangerfjord, Nordfjord, Møre og Romsdal, fjord formation, bolide impact, mega-tsunami, Younger Dryas impact hypothesis, hydraulic plucking, bedrock erosion, Charlie-Gibbs Fracture Zone, Norwegian continental shelf, Mjølnir crater, Silverpit crater, deep-water emergence, impact geomorphology, impact science.

Table of Contents

Introduction

This paper originated in research on the Møre og Romsdal region of western Norway. Two features of the region's geology prompted further investigation: the Sunnmøre-Nordfjord area experienced significantly less Younger Dryas ice readvance than regions both to the north and south despite equivalent precipitation (Mangerud et al., 1979; Larsen et al., 1984), and the region contains Norway's deepest and most dramatic fjords — features conventionally attributed to progressive glacial erosion across multiple Quaternary cycles (Gregory, 1913; Nesje et al., 1992; Holtedahl, 1967).

Additional findings during the investigation of these anomalies led to the hypothesis that the Møre og Romsdal fjords were carved by catastrophic hydraulic focusing of trans-Atlantic bolide-generated mega-tsunami at approximately 12,800 BP — the onset of the Younger Dryas. The physical model draws on the megaflood geomorphology literature, which has demonstrated that catastrophic hydraulic events produce landscape-scale bedrock erosion in geologically instantaneous timescales (Gupta et al., 2007; Lamb & Fonstad, 2010; Baynes et al., 2015), with floods responsible for >2.5 km of canyon retreat in Iceland shown to be only a third as large as previously estimated (Stucky de Quay et al., 2021).

This paper draws on the dam spillway engineering literature — largely uncited in geomorphology — which documents bedrock scour of tens of meters per event at flow velocities of 15–25 m/s, including removal of 1,200-tonne boulders in a single flood (Stratford et al., 2013) and 90 m of cumulative bedrock scour below the Kariba Dam (Bollaert, 2004; Noret et al., 2012).

The identification of the Boston Basin as a candidate YD-age bolide swarm impact site (Gjøvik, 2026) — supported by multiple lines of evidence including: fault-controlled kaolinization to depths exceeding 300 feet (Kaye, 1967), rock recrystallization at 175–250°C with no heat source for 400 million years (Anderson, 2008), fresh olivine that cannot persist at surface conditions without recent emplacement (Rahm, 1962), a chondritic mineral assemblage (Thompson, 2020; Ross, 1990), and elevated platinum group elements (Tuit, 2000) — provides the source location, impactor composition, and energy transfer geometry.

The over 200 drumlins in the Boston Basin (La Forge, 1932), reinterpreted by Gjøvik (2026) as ejecta mounds from individual swarm fragments, constrain the swarm size and the wave generation calculation. The Charlie-Gibbs Fracture Zone — the deepest passage through the Mid-Atlantic Ridge — provides the trans-Atlantic energy pathway.

After developing the Møre og Romsdal model, the author examined whether the mechanism could apply more broadly. Norwegian fjords do not share a single orientation, and found that each orientation group faces a water body containing a confirmed impact structure — with fjord morphology correlating to impact age despite uniform bedrock lithology across regions. Examination of the global record revealed that, it appears, every major fjord coastline faces an ocean basin with documented impact events.

These findings suggest that bolide-generated mega-tsunami may be the primary mechanism for fjord formation worldwide, with glacial processes modifying rather than creating the channels. This paper presents the western Norway case study in detail and the broader pattern in outline, and calls for researchers at other fjord coastlines to evaluate the mechanism independently.

This research indicates that glacial striations on upper fjord walls record later glacial occupation of channels that already existed, not the formation of those channels. The model is consistent with a sequence of tectonic fracture valleys, catastrophic excavation, and subsequent glacial modification.

Study Area

The Western Norwegian Fjord Zone

The study area encompasses the western Norwegian fjord coast between approximately 59°N and 63°N — from the Rogaland coast south of Stavanger to the northern boundary of Møre og Romsdal county. This zone contains all of Norway's deepest and most morphologically dramatic fjords, including the world's second-deepest (Sognefjord) and three UNESCO World Heritage fjords (Geirangerfjord, Nærøyfjord). Aarseth (1997) estimated that approximately 150 km^3 of sediments are present in the western fjords between 59–63°N, with 90% of this material deposited during the deglaciation of the last Weichselian ice sheet. Less than 10% of the fjord sediments in this zone predate the Last Glacial Maximum.

The coast faces the open North Atlantic with no significant shielding between the Norwegian shelf edge and the Mid-Atlantic Ridge. The Charlie-Gibbs Fracture Zone — the deepest trans-Atlantic passage through the Mid-Atlantic Ridge — exits onto the eastern Atlantic basin along a great-circle path toward this coastline. The continental shelf is narrow, typically 50–150 km wide, dropping from less than 200 m depth at the shelf edge to the deep Norwegian Sea basin.

The bedrock across the study area is predominantly Precambrian gneiss of the Western Gneiss Region, with Caledonian nappes in the inner fjord areas and Devonian sedimentary basins exposed between Sognefjord and Nordfjord (Sigmond et al., 1984; Nesje & Whillans, 1994). The dominant structural grain of the Caledonian orogen strikes northeast-southwest, oblique to the east-west trend of most major fjords.

The Primary Case Study

The Western Norwegian coast, at the northern end of the study area (62–63°N), is the most directly exposed to the open Atlantic and contains the paper's primary case study. The dominant feature is the Storfjord system — Norway's fifth-longest fjord at 110 km — which penetrates inland from the outer Sunnmøre coast near Hareid eastward to the villages of Tafjord and Geiranger.

At the village of Stranda, Storfjord branches into two major arms: Sunnylvsfjorden (26 km) extending into the UNESCO-listed Geirangerfjord (15 km, maximum depth ~260 m), and Norddalsfjorden extending into Tafjorden. Hjørundfjord (20 km), an arm off the main Storfjord, is surrounded by the Sunnmørsalpene mountain range with peaks reaching 1,700 m directly from the fjord surface (Suul et al., 2004). The fjord walls are 1–2 km wide with cliffs reaching 1,300 m in places.

Romsdalsfjord, in the Romsdal district to the north, extends 88 km with a maximum depth of approximately 800 m. Unlike most western fjords, Romsdalsfjord has several islands and a wider outer section before narrowing inland. The adjacent mountains include Trollveggen — the highest vertical rock face in northern Europe at approximately 1,000 m — and peaks exceeding 1,900 m in the Romsdalsfjella range.

The MOR coast is the location of the Sunnmøre-Nordfjord anomaly: this stretch experienced significantly less Younger Dryas ice readvance than regions to both the north (Trøndelag) and south (Bergen–Sognefjord), despite equivalent precipitation and ice accumulation area geometry (Mangerud et al., 1979; Larsen et al., 1984). YD end moraines have been documented between Hardangerfjord and Sognefjord (Aarseth & Mangerud, 1974), but are absent or minimal across the Sunnmøre-Nordfjord section. Under the catastrophic model, this anomaly reflects the mega-tsunami's destruction of the YD ice front in the most Atlantic-exposed section.

The Southern Fjords: Vestland and Rogaland

South of MOR, the study area includes progressively deeper fjords. Nordfjord (105 km, 565 m) lies at the MOR-Vestland boundary. Sognefjord — Norway's longest at 204 km and deepest at 1,308 m (bedrock at approximately 1,500 m below sea level; Nesje et al., 1992) — dominates the Vestland coast at approximately 61°N. The fjord averages less than 5 km width and exceeds 1,000 m depth for approximately 100 km of its length (Nesje & Whillans, 1994). Sognefjord's east-west axis bears no clear relation to the northeast-southwest Caledonian fold pattern through which it cuts (Nicholson, 1963; Nesje & Whillans, 1994).

Hardangerfjord (179 km, approximately 860 m; Fossen & Hurich, 2005) follows the Hardangerfjord Shear Zone — a tectonic structure that provides the structural template the catastrophic model requires. Further south, Lysefjord (42 km, approximately 460 m) near Stavanger represents the southernmost of the deep western fjords and the least depth of the group — consistent with its more sheltered, southward-facing orientation.

The depth gradient across the study area is consistent with the catastrophic model: depth correlates with Atlantic exposure rather than with bedrock type, which is uniform gneiss throughout. Sognefjord (1,308 m) faces most directly west; Hardangerfjord (860 m) trends more to the northwest; Lysefjord (460 m) faces southwest with partial island shielding. This gradient is not predicted by the glacial model, which attributes depth to ice thickness and velocity — factors that do not correlate with ocean exposure direction.

Multiple Events

The western Norwegian coast faces the open Atlantic, which has experienced multiple confirmed and candidate bolide impacts over geological time.

While this paper develops the MOR physical model specifically through the Younger Dryas event at ~12,800 BP, it is possible that some western fjords record earlier Atlantic impact events. No direct geochronological dating of fjord excavation exists — cosmogenic exposure dating of submerged fjord walls has not been attempted, and the fjord sediment record captures only the post-excavation infill (Aarseth, 1997). The observation that 90% of fjord sediments were deposited during the last deglaciation

(Aarseth, 1997) is consistent with either progressive glacial erosion followed by deglacial infill, or catastrophic excavation at ~12,800 BP followed by the same deglacial infill. These interpretations cannot currently be distinguished without direct dating of the bedrock erosion surface beneath the sediments.

The YD readvance pattern provides a possible discriminator. The absence of YD moraines in the Sunnmøre-Nordfjord section — the most Atlantic-exposed stretch — is specifically predicted by the YDI model and is difficult to explain under the glacial model. The presence of YD moraines between Hardangerfjord and Sognefjord (Aarseth & Mangerud, 1974) indicates that the YD ice front was intact in those areas, which could mean either that the mega-tsunami had less impact further south (consistent with reduced wave energy at the margins of the impact zone) or that those fjords already existed from an earlier event and the YD ice was able to readvance into pre-existing channels. This paper presents this as an open question that future research — particularly cosmogenic dating of exposed bedrock surfaces within different fjords — could resolve.

Anomalies in the Standard Glacial Model

The Distribution Anomaly

Fjords concentrate on west-facing, ocean-exposed coasts globally: Norway's western seaboard, Chile's Pacific coast, New Zealand's Fiordland, Alaska, British Columbia, and Greenland's ocean-facing margins.

The standard explanation holds that fjords form where fast-flowing outlet glaciers drain through topographically confined channels to the sea (Glasser et al., 2008). This accounts for why fjords require both mountainous terrain and coastal proximity, but it does not fully explain the directional selectivity. Sweden, which was covered by the thickest portion of the Fennoscandian ice sheet during most glacial periods, has virtually no fjords. The standard response — that Swedish ice flowed away from the Atlantic margin rather than toward steep coastal topography — is partially satisfactory, but Scotland presents a case worth further investigation: fjords (sea lochs) are found almost exclusively on the west coast, despite glacial outlet activity on both coasts. Whether the east coast had comparable topographic confinement and ice discharge rates requires detailed comparison, but the directional asymmetry is suggestive.

This directional pattern — Atlantic-facing coasts with deep fjords, non-Atlantic-facing coasts without — is consistent with this proposed model, in which the deepening agent propagated from the Atlantic. It is less naturally explained by glacial processes, which should not exhibit systematic compass-direction dependence independent of ice flow geometry.

The Depth Anomaly

Norwegian fjords achieve extraordinary depths in crystalline metamorphic bedrock. Sognefjord's bedrock floor lies approximately 1,500 m below sea level, with an estimated 2,000 km^3 of rock removed (Nesje et al., 1992).

The strongest version of the glacial erosion argument invokes quarrying (block plucking by ice), not abrasion, as the primary mechanism. Quarrying rates can exceed abrasion rates by an order of magnitude, reaching several mm/yr under favorable conditions (Hallet et al., 1996; Herman et al., 2015). At 5 mm/yr, the full depth of Sognefjord could theoretically be excavated in approximately 300,000 years — within the Quaternary timeframe.

However, sustained quarrying at maximum rates requires specific conditions: warm-based ice, high basal water pressures, and abundant meltwater — conditions that do not persist uniformly across glacial cycles. The effective time-averaged rate is likely substantially lower than the peak rate. Moreover, the question of why Norwegian fjords achieved such extreme depths while comparably glaciated settings with similar bedrock (e.g., interior Scandinavia) did not, remains open.

The Dating Gap & Sill Problem

No published study has directly dated the progressive deepening of Norwegian fjord walls across multiple glacial cycles using cosmogenic nuclide exposure dating (^{10}Be, ^{26}Al). The Quaternary timeframe for fjord carving is inferred from erosion rate extrapolation and geomorphological analogy, not measured directly. Nesje et al. (1992) acknowledged that the preglacial valley floor cannot be accurately reconstructed along the present fjord.

Fjord sills — shallow ridges at fjord mouths — are conventionally interpreted as terminal moraines. However, sills would have impeded glacial flow and reduced erosive capacity at the fjord mouth, while the deepest excavation occurs farther inland. This tension has been noted in the literature (e.g., the discussion in Nesje et al., 1992) but not fully resolved.

Offshore Sediment Deposits

If Norwegian fjords were carved progressively across multiple glacial cycles, the excavated material — estimated at ~2,000 km³ for Sognefjord alone (Nesje et al., 1992), and tens of thousands of km³ across all Norwegian fjords — was transported offshore incrementally over approximately 2.8 million years. If instead a catastrophic event excavated a significant fraction of this volume in a single episode, the resulting deposit would be concentrated at a single stratigraphic horizon.

The Norwegian continental shelf between 61°N and 68°N is among the most thoroughly surveyed margins on Earth. During the last 2.8 million years, the shelf prograded up to 150 km westward, depositing over 1,000 m of Quaternary sediments in the Naust Formation, mapped extensively with 3D seismic data, close to 200 gravity cores, and petroleum exploration wells (Dahlgren et al., 2002; Ottesen et al., 2009). Large-volume events leave unmistakable signatures in this record: the Storegga Slide deposited approximately 3,500–5,580 km³ and is readily identifiable across multiple core sites (Bondevik et al., 2005; Bugge et al., 1988).

The sediment record at the Younger Dryas horizon has been characterized, but exclusively within a glaciomarine interpretive framework. Investigation of close to 200 gravity cores from the northern Norwegian shelf describes the ~13,000 BP horizon as a transition from pebbly pelite deposition in troughs and iceberg ploughing on banks to "a period with incipient winnowing" on the deep banks and "deposition of sandy pelite" in the troughs (Vorren et al., 1984). In the Nordfjord system — within the Sunnmøre-Nordfjord area central to this model — up to 360 m of sediments were deposited during the last deglaciation, with the greatest volume accumulating during the Allerød recession, the Younger Dryas readvance, and the subsequent retreat (Hjelstuen et al., 2009). "Extensive re-sedimentation" is documented in the late Younger Dryas and succeeding approximately 1,000 years (Hjelstuen et al., 2009). Early investigations of Norwegian Sea bottom cores recovered "a gray, unsorted, sandy and pebbly clay" containing displaced benthonic foraminifera from shallower depths, extending to approximately 900 m water depth (Holtedahl, 1959). These deposits have been uniformly classified as glaciomarine.

Whether any portion of the documented resedimentation at the YD horizon represents catastrophic excavation rather than glaciomarine processes is an open question that existing data can address. No published study has examined Norwegian shelf sediments at the ~12,800 BP horizon using established tsunami deposit diagnostic criteria — specifically, chaotic mixing of organisms from multiple depth zones, poorly-sorted mixed lithologies inconsistent with local bedrock provenance, and sedimentary structures diagnostic of high-energy emplacement (Mamo et al., 2009; Pilarczyk et al., 2020). The Norwegian Geological Survey (NGU) and petroleum operators hold core material spanning this horizon that could be reanalyzed with these criteria without new fieldwork.

The author notes that the absence of a recognized catastrophic deposit in the existing literature does not constitute evidence against the model, because the question has not been asked. Conversely, if targeted reanalysis demonstrates that YD-horizon sediments on the Norwegian shelf are compositionally and structurally consistent with glaciomarine deposition and inconsistent with catastrophic emplacement, this would constitute significant evidence against the catastrophic model.

A Trans-Atlantic Catastrophic Event

The Younger Dryas Impact Hypothesis

Firestone et al. (2007) proposed that a cosmic impact at approximately 12,800 cal BP triggered the Younger Dryas cooling. Evidence includes platinum anomalies at 26+ YDB sites including the GISP2 ice core; microspherules at 50+ sites on four continents, independently replicated by multiple groups; shock-fractured quartz at eastern US sites consistent with airburst events (Moore et al., 2024); impact proxies including cometary dust and platinum anomalies in marine sediment cores from Baffin Bay — the first marine evidence for the YDIH (Moore et al., 2025); and biomass burning consuming approximately 10 million km² (~9% of terrestrial biomass) — the largest episode in 120,000 years of ice core records (Wolbach et al., 2018a, 2018b). Bayesian chronological analysis confirmed synchronous age across four continents (Kennett et al., 2015).

The hypothesis remains debated. Holliday et al. (2023) published a comprehensive critique arguing that the evidence involves flawed methodologies and irreproducible observations. Key concerns include the reproducibility of nanodiamond identification, the terrestrial (rather than extraterrestrial) origin of some putative impact markers, and the absence of a dated impact crater. A separate study reporting shocked quartz at Clovis sites (Kennett et al., *PLOS One*, 2025) was retracted in February 2026 over concerns about the underlying age model and sampling strategy, though the shocked quartz findings reported independently by Moore et al. (2024) at different sites using different methods are unaffected. These concerns are substantive and should not be dismissed. However, Sweatman et al. (2024a, 2024b) and Powell (2022) have rebutted several specific claims in Holliday et al. and argued that the critique does not constitute falsification of a hypothesis supported by evidence at more than 50 sites.

Of particular relevance to this paper, the Boston Basin presents a documented explanatory vacuum in its own geological literature. Kaye (1984) wrote that Boston would "probably be one of the last to be mapped satisfactorily." Barosh (2011) described the surficial deposits as "probably the most complex in the country" and a "bewildering array of strata that may change abruptly over a short distance." No consensus explanation exists for the basin's geology, despite over a century of study. Critically, the Lexington Substage — the mass debris event that shaped the basin's surficial geology — is dated to approximately 12,900 BP by standard regional stratigraphy (Barosh, 2011), but Barosh also notes the Lexington Substage readvance "only extended into valleys north and west of Boston" — meaning the features attributed to glacial activity in Boston proper formed without glaciers being present in Boston at that time.

The mineral and thermal data from the basin compound this problem. Anderson (2008) documented rock recrystallization at 175–250°C across the argillite formation, with no documented heat source in the basin for over 400 million years. Kaye (1967) documented fault-controlled kaolinization to depths exceeding 300 feet — a hydrothermal signature — but could not identify the heat source. Rahm (1962) found fresh olivine alongside partially altered chlorite and magnetite in the same rock body; olivine is among the least stable rock-forming minerals at surface conditions and cannot persist for 400 Ma without a recent emplacement event. The argillite further contains a chondrite-compatible mineral assemblage including albitic feldspar (Ab_{84}), ilmenite in three alteration stages, chromite with elevated Cr and Ni, and platinum group elements at five times crustal background (Thompson, 2020; Ross, 1990; Tuit, 2000; see Gjøvik, 2026, for the complete mineral analysis).

Gjøvik (2026) proposed that the most parsimonious explanation for this suite of anomalies is a bolide swarm impact at approximately 12,900 BP. For this paper, two constraints from that identification are directly relevant: the chondritic mineral signature indicates an impactor density of approximately 3,400 kg/m³ — 3.4× the cometary density assumed in early YDIH literature — which affects the energy calculations; and the basin's offshore extension into Massachusetts Bay means some swarm fragments impacted shallow continental shelf directly, partially resolving the wave generation uncertainty.

This model does not require the YDIH to be fully validated. It requires that a catastrophic event of sufficient energy occurred in the North Atlantic region at approximately 12,800 BP. The thermohaline disruption triggering the Younger Dryas cooling is not controversial; its cause remains an open question to which the YDIH is one candidate answer.

Geographic Alignment

The proposed impact zone — the Laurentide ice sheet over northeastern North America — is approximately 5,500 km across open Atlantic from the Norwegian west coast. The ice sheet's terminal moraines in New England (Long Island, Cape Cod, Martha's Vineyard) mark the approximate margin, with ice thickness of 1.5–3 km in this region at 12,800 BP (Balter-Kennedy et al., 2024).

If the Boston Basin identification (Gjøvik, 2026) is correct, the source geometry becomes specific. The Boston Basin is a ~50-km-long, east-tilting asymmetrical rift basin whose eastern side is submerged by Boston Harbor and Massachusetts Bay, extending offshore beneath Stellwagen Bank (Barosh, 2011; Billings, 1979). This places the impact zone at the margin of the New England continental shelf — a ~200-km-wide platform at ~50 m depth at 12,800 BP sea levels. Bolide swarm fragments impacting this geometry would have struck both the ice sheet and the shallow marine environment at the shelf margin. Fragments impacting water generate tsunamis with substantially higher efficiency than ice-sheet impacts, and the shelf itself acts as a waveguide for the resulting wave energy.

Deglaciation State and the Younger Dryas Readvance Anomaly

A critical prerequisite for this model is that the Norwegian fjord channels were accessible to oceanic energy at the time of the proposed event. The deglaciation timeline for western Norway constrains this — and reveals an anomaly that this model may explain.

In the Møre og Romsdal region specifically, Sollid and Sørbel (1979) reconstructed ice margins during the Younger Dryas and found that "the glacier front lay at the fjord heads" with the outer and middle portions of the fjords ice-free. They noted that "raised beach features indicate that the deep fjords became ice-free at an early stage due to calving." Mangerud et al. (1979) dated the initial coastal deglaciation in the Bergen area to ~12,600 BP and in Nordfjord to slightly before 12,300 BP. The outer coast was thus largely deglaciated at or before the time of the proposed event (~12,800 BP).

During the YD cold period, the Scandinavian ice sheet readvanced along most of the Norwegian coast. In the Bergen/Hardangerfjorden area, this readvance was substantial: Mangerud et al. (2016) document a ~40–50 km readvance into fjords that had been ice-free during the Allerød, supported by 90 radiocarbon dates from 36 overrun sites. YD moraines are mapped more or less continuously around the Scandinavian ice sheet (Andersen et al., 1995). Yet in the Sunnmøre-Nordfjord area of Møre og Romsdal, Mangerud et al. (1979) found "no large re-advance of the inland ice occurred during the Younger Dryas" — only local glacier formation at higher elevations.

This is anomalous. The region with the most direct Atlantic exposure — precisely where this model predicts maximum tsunami energy delivery — shows significantly less YD readvance than regions both to the north and south. Direct bolide impacts on Norway can be effectively ruled out as an explanation: Norway has three confirmed impact structures (Mjølnir, Gardnos, Ritland), all dating to hundreds of millions of years ago, and active crater-identification programs at the University of Oslo have not identified any late Pleistocene candidates. No YDB impact proxies (microspherules, shocked quartz, meltglass, platinum anomalies) have been reported from Norwegian terrestrial sites.

The regional difference is not easily attributable to differential precipitation: Larsen et al. (1984) found that the heaviest YD precipitation in Norway was in the Bergen–Nordfjord region, with increasing continentality both north and south. Bergen and Sunnmøre-Nordfjord received comparable or equivalent precipitation, yet their readvance histories diverge dramatically. Conventional explanations might invoke differences in bed topography, calving dynamics, or accumulation area geometry, but no specific mechanism has been proposed in the literature to account for why the Sunnmøre-Nordfjord region, situated in the same high-precipitation zone, did not experience a comparable inland ice readvance.

This model offers a specific explanation: if the trans-Atlantic tsunami catastrophically destabilized coastal ice in the most exposed region at ~12,800 BP (the very onset of the YD), the ice sheet would have had to rebuild from a position further inland than in adjacent, less exposed regions. The mechanism for tsunami-ice interaction has empirical support: MacAyeal et al. (2006) documented Antarctic ice calving linked to transoceanic wave propagation from the 2004 Indian Ocean tsunami, and Brunt et al. (2011) documented similar calving triggered by the 2011 Tōhoku tsunami — both at far lower energies than those proposed here. A tsunami arriving at a tidewater glacier terminus would cause calving and mechanical destabilization of grounded ice, potentially removing ice from the outer and middle fjord system.

Ice removal consumes energy that is then unavailable for bedrock erosion. The energy budget must be partitioned between ice destabilization and bedrock plucking, and this partitioning has not been quantified. However, a multi-pulse scenario (multiple swarm fragments over hours) would deliver sequential waves: earlier pulses destabilize and remove ice, later pulses act on exposed or partially exposed bedrock. The ~1,200-year YD cold period may not have been sufficient for the ice to readvance the full distance to the coast from this setback position.

Catastrophic Hydraulic Focusing: The Physical Model

Energy Scale

For individual fragments at 20 km/s, the kinetic energy depends critically on impactor density. Early YDIH literature assumed cometary density (~1,000 kg/m^3). However, the chondritic mineral fingerprint documented at the proposed Boston Basin impact site (Gjøvik, 2026) indicates ordinary chondrite composition with a density of approximately 3,400 kg/m^3. Both scenarios are prsented:

Fragment	Cometary (1,000 kg/m^3)	Chondritic (3,400 kg/m^3)
500-m	1.3×10^{19} J	4.5×10^{19} J
1-km	1.05×10^{20} J	3.6×10^{20} J
2-km	8.4×10^{20} J	2.9×10^{21} J

For comparison, the 2004 Indian Ocean tsunami released ~1.1×10^{17} J. Even a single cometary 500-m fragment represents approximately 100× the Indian Ocean tsunami energy; a chondritic fragment of the same size represents ~400×. Multiple fragments, as the swarm hypothesis proposes, compound these values.

The proxy evidence independently constrains the minimum event energy. Biomass burning consuming 9% of Earth's terrestrial surface requires atmospheric energy injection well above 10^{18} J (Wolbach et al., 2018a).

Ice Sheet Impact, Partial Marine Impact, and Wave Generation

The YDIH proposes impact on a 2–3 km continental ice sheet — a scenario that has never been computationally modeled for wave generation. This has been the single most important uncertainty in this model. The published 1% kinetic-energy-to-wave conversion figure (Gisler et al., 2018) was derived for deep-ocean impacts. Ice sheet impacts involve qualitatively different energy pathways: explosive ice vaporization, atmospheric blast coupling, seismic ground coupling, and catastrophic meltwater release.

However, if the Boston Basin identification (Gjøvik, 2026) is correct, the impact was not purely an ice-sheet event. The basin extends offshore into Massachusetts Bay, and a bolide swarm distributed across a ~50 km zone would have included fragments impacting the shallow marine environment at the continental shelf margin, either directly or through ice that was thinner or absent over the marine portions of the basin. Fragments impacting shallow water generate tsunamis through direct displacement — a well-characterized process for which the standard 1% conversion efficiency applies, and shallow-water impacts can exceed this figure. The wave generation uncertainty thus applies primarily to the ice-sheet-impacting fraction of the swarm, not to the entire event.

Additionally, a distributed swarm across ~50 km generates a more planar wavefront than the point-source geometry assumed in the Ward (2003) model used for calibration. Planar waves attenuate more slowly with distance than radially spreading point-source waves ($1/\sqrt{r}$ vs. $1/r$ in two dimensions). The Ward-calibrated estimates may therefore underestimate arrival amplitudes for a distributed source.

A lower bound can be established using the Krakatoa 1883 eruption as an analog for atmospheric-blast-to-wave coupling. Krakatoa released approximately 200 megatons TNT equivalent (~8.4×10^{17} J). Its atmospheric blast wave circled the globe multiple times and generated sea level disturbances of approximately 1 m at distances exceeding 5,000 km through atmospheric coupling alone — not through direct water displacement (Harkrider & Press, 1967; Pararas-Carayannis, 2003). Even a single 500-m chondritic fragment represents approximately 10,700 MT — roughly 53× Krakatoa's energy.

Atmospheric-ocean coupling was bound using the Krakatoa analog. Standard atmospheric blast scaling (Sachs scaling) gives overpressure at fixed distance proportional to E^(1/3). However, ocean surface displacement from atmospheric pressure waves depends on impulse — the product of pressure and duration — rather than peak pressure alone. Since positive-phase duration also increases with source energy (approximately as E^(1/6) from Sachs scaling), total impulse scales as E^(1/2). Scaling as the square root of the energy ratio therefore represents the impulse-dependent upper bound, while E^(1/3) represents the peak-pressure lower bound. For a single 500-m chondritic fragment (~53× Krakatoa's energy), these bounds yield approximately 4–7 m waves at 5,000 km from atmospheric coupling alone.

A 1-km chondritic fragment (~424× Krakatoa) yields approximately 8–21 m. These figures represent only the atmospheric pathway; seismic ground coupling, direct mechanical coupling at the ice-ocean margin, direct water-impact wave generation, and catastrophic meltwater release would add additional wave energy through independent mechanisms.

The subsequent calculations as presented as conditional: *if* the combined conversion efficiency produces arrival waves in the range suggested by these bounding estimates and the Ward (2003) calibration, *then* the following erosion

Continental Shelf Geometry

The energy transfer geometry of the proposed event is significant. The Boston Basin sits at the western margin of the New England continental shelf — approximately 200 km wide and only ~50 m deep at 12,800 BP sea levels (Gjøvik, 2026; Barosh, 2011). The basin's eastern extension into Massachusetts Bay means that impact-generated meltwater, displaced material, and waves from marine-impacting swarm fragments enter the shelf directly at its inner margin. Shallow-water waves propagate without dispersion (wave speed ~22 m/s at 50 m depth), maintaining coherence across the full shelf width. Upon reaching the shelf edge (depth transition to ~3,000 m), the wave would acquire the long wavelength characteristic of efficient trans-oceanic propagation.

This geometry likely produces more efficient wave generation than a comparable deep-ocean impact, because the shallow shelf constrains the wave energy vertically, preventing the deep-water radial spreading that dissipates energy from open-ocean impacts. Quantifying this effect requires computational modeling, but the qualitative implication is that Ward's (2003) deep-ocean impact model, used for calibration, may underestimate wave amplitudes for a shelf-margin source.

Mid-Atlantic Ridge Barrier and the Charlie-Gibbs Fracture Zone

Trans-Atlantic wave propagation from a New England source to the Norwegian coast must cross the Mid-Atlantic Ridge (MAR), which rises to approximately 2,000 m depth or shallower along much of its length and constitutes a significant topographic barrier to deep-water wave propagation. However, the MAR is interrupted by transform faults and fracture zones that provide deep-water passages.

The Charlie-Gibbs Fracture Zone (CGFZ), at approximately 52°N, is the most prominent interruption of the MAR between the Azores and Iceland — a 340-km offset comprising two parallel transform faults extending approximately 2,000 km in length, with depths ranging from 700 to 4,500 m (Heezen & Ewing, 1963). It is described as one of the deepest connections between the northwest and northeast Atlantic. The CGFZ lies directly in the great-circle propagation path from the New England coast to the Møre og Romsdal region of Norway.

This geometry has specific implications for wave energy distribution. Wave energy propagating eastward from a New England source would be partially blocked or attenuated by the MAR along most of its north-south extent. The degree of blockage depends on the ratio of wave wavelength to ridge width: very long wavelength waves (100+ km) would be partially transmitted over the ~2,000 m deep ridge crest with reduced amplitude, while shorter wavelength components would be more effectively blocked. The CGFZ would provide a preferential pathway through which wave energy of all wavelengths passes to the eastern Atlantic with less attenuation. Energy exiting the eastern end of the CGFZ at approximately 52°N would then spread into the eastern Atlantic basin, arriving with preferentially concentrated — though not exclusively focused — intensity along the Norwegian coast between approximately 60°N and 64°N, the Møre og Romsdal region where the deepest and most dramatic fjords are found.

This observation upgrades the directional distribution prediction from a simple compass-bearing correlation to a specific submarine-topography prediction: computational wave propagation modeling using realistic Atlantic bathymetry should show energy concentration at the Møre og Romsdal coast, not merely at all west-facing coastlines. The New England Seamount Chain — over 30 submarine peaks extending 1,000+ km from Georges Bank — would also interact with the outgoing wave field and must be included in any realistic propagation model.

Trans-Atlantic Propagation

The model was calibrated against Ward (2003), who modeled a 1.1-km asteroid impact 600 km off the US east coast (KE ~2.4×10^{20} J) and found 20–23 m waves reaching Ireland, 16 m in Northern Europe. Kharif and Pelinovsky (2005) obtained consistent results.

The model was scaled from Ward's results using the energy ratio. Initial tsunami amplitude is bounded by the depth of the transient impact cavity, which scales with impactor radius through Schmidt-Holsapple crater scaling laws (Schmidt & Holsapple, 1982; Ward & Asphaug, 2000). Since cavity depth scales approximately as R_I^0.78 while kinetic energy scales as R_I^3, wave height scales approximately as E^0.25 — a relationship derived from crater physics, not wave propagation, and robust across impactor sizes in the Ward & Asphaug (2000, 2003) framework.

Both cometary and chondritic density scenarios are presented, and do not apply additional correction factors for the shelf waveguide or distributed-source geometry discussed, both of which would increase these estimates:

Scenario	**Density**	**Total KE (J)**	**vs. Ward**	**Est. arrival at Norway (m)**
Single 500-m	Cometary	1.3×10^{19}	0.05×	~10
Single 500-m	Chondritic	4.5×10^{19}	0.19×	~14
Three 500-m	Cometary	3.9×10^{19}	0.16×	~13
Three 500-m	Chondritic	1.35×10^{20}	0.56×	~18
Single 1-km	Cometary	1.05×10^{20}	0.43×	~17
Single 1-km	Chondritic	3.6×10^{20}	1.5×	~23
2-km equivalent	Cometary	8.4×10^{20}	3.5×	~28
2-km equivalent	Chondritic	2.9×10^{21}	12×	~38

*Note: The atmospheric coupling estimates (Section 4.2) use impulse-dependent E^(1/2) scaling. If peak-pressure-dependent E^(1/3) scaling applies instead, these estimates reduce by approximately 40–50%, yielding 4–12 m rather than 7–21 m for the atmospheric component. Even at the lower bound, the atmospheric pathway contributes meaningful wave energy independent of the direct-impact pathway.

These estimates are conservative: they assume point-source radial spreading (a distributed swarm source would attenuate more slowly), do not account for the potentially enhanced wave generation efficiency of partial marine impact or shelf waveguide effects. The Krakatoa bounding estimates suggest that atmospheric coupling alone could contribute an additional 7–21 m for chondritic fragments, independent of these figures.

Fjord Amplification

Wave amplification in narrowing channels follows Green's Law, empirically validated for Norwegian fjords by Bondevik et al. (2005) in modeling the Storegga tsunami. Vasskog et al. (2013) modeled Storegga amplification to approximately 40 m in inner Geiranger from approximately 10 m at the outer coast. For typical Møre og Romsdal geometry (mouth ~5 km to inner ~500 m width; 200 m to 50 m depth), the amplification factor is approximately 4.5×.

At the wave heights and amplification factors involved, wave breaking would likely occur before full Green's Law amplification is achieved. Breaking dissipates organized wave energy into turbulence. However, breaking waves in confined channels concentrate turbulent energy at the bed and walls, potentially increasing rather than decreasing erosive capacity at those surfaces. The Green's Law heights are presented as upper bounds on organized wave height, while noting that the total energy delivery to the channel boundaries may be comparable or greater due to turbulent concentration.

Conditional estimates for inner fjord conditions:

Arrival (m)	Green's Law height (m)*	Est. flow velocity (m/s)**
10	45	20
17	76	34
28	126	56
39 (with atm. coupling)	175	78

*Upper bound on organized wave height; wave breaking at these amplitudes would convert organized wave energy to turbulence concentrated at channel boundaries. **Velocity estimates assume shallow-water wave dynamics; turbulent bore velocities may differ. Dam-break bore theory (the Ritter solution and extensions) suggests that bore-front velocities in confined channels scale as approximately $2\sqrt{(gh)}$ where h is the bore depth, which for the energy scales involved here yields velocities of similar order to the shallow-water estimates. Detailed bore propagation modeling in fjord geometry would refine these figures.

Erosion Mechanisms, Rates, and Analogs

Plucking. Whipple et al. (2000) established that plucking dominates bedrock erosion wherever rocks are well jointed on a submeter scale, and that above a threshold shear stress, erosion rates increase approximately linearly with stress. For 1-m blocks in jointed metamorphic rock, the critical flow velocity is approximately 5–10 m/s (Baynes et al., 2015; Lamb et al., 2014). The modeled velocities (20–78 m/s) exceed this by factors of 2–16×. At these velocities, shear stress — which scales as the square of velocity — exceeds the plucking threshold by factors of 4–250×.

Cavitation. Onset occurs at >10–15 m/s in confined channels (Whipple et al., 2000). These velocities substantially exceed this threshold. Cavitation produces extreme localized pressure fluctuations within joint networks that contribute to block loosening (Bollaert & Schleiss, 2005).

Dynamic pressure propagation. The dam spillway engineering literature provides a critical additional mechanism not addressed in the megaflood geomorphology literature. Bollaert and Schleiss (2005) demonstrated that high-velocity flow generates pressure fluctuations that propagate into bedrock joint networks ahead of the erosion front. Pan et al. (2014) showed that even when individual pressure cycles do not exceed the plucking threshold, cumulative irreversible displacement of blocks subjected to repeated pressure fluctuations can gradually cause weak planes to become fully persistent, leading to plucking. In a multi-pulse tsunami scenario, earlier pulses that do not fully dislodge blocks at depth may weaken joint surfaces and prepare them for removal by subsequent pulses — a progressive failure mechanism that increases the effective erosion depth per event sequence beyond what any single pulse would achieve.

Catastrophic Erosion Analogs

The megaflood analogs cited in geomorphology literature demonstrate landscape-scale erosion in jointed bedrock:

The English Channel megaflood. Gupta et al. (2007, Nature) identified a bedrock valley up to 50 m deep and tens of km wide on the Channel floor. Gupta et al. (2017, Nature Communications) confirmed two-stage catastrophic breaching.

Channeled Scablands. The Missoula Floods carved deep channels in basalt across 16,000+ km² at peak discharges of 10–17 Sv (Baker, 1973, 2009).

Iceland canyons. Baynes et al. (2015, PNAS) documented >2 km knickpoint retreat and ~100 m incision through jointed basalt during individual extreme flood events. Critically, a subsequent study (Stucky de Quay et al., 2021, Communications Earth & Environment) demonstrated that the floods responsible were at most a third of previous discharge estimates, yet still achieved >2.5 km of knickpoint retreat — leading the authors to conclude that "the extent of bedrock erosion is not necessarily controlled by discharge" and that jointed volcanic flows "allow shear and drag from flood waters to topple basalt columns with relative ease." This finding directly addresses the transient-versus-sustained flow question: erosive efficiency per unit discharge is higher than models predict.

However, the megaflood analogs involve flow velocities substantially below those proposed in this model. A more directly relevant velocity regime is documented in the dam spillway engineering literature, where high-velocity flow over jointed bedrock produces erosion that dam engineers describe as exceeding model predictions.

Dam spillway scour. The Kariba Dam on the Zambezi River (128 m height) has scoured bedrock to approximately 90 m below the tailwater level through repeated spillway operation (Bollaert, 2004; Noret et al., 2012). The Wivenhoe Dam in Australia (2011 flood) eroded boulders weighing up to 1,200 tonnes from its plunge pool at flow velocities less than 25 m/s and head differentials rarely exceeding 30 m — erosion that spillway engineers described as "almost unbelievable" and that exceeded existing predictive models (Stratford et al., 2013). The Copeton embankment dam (113 m) in Australia experienced significant water flow that created a 20 m deep gorge in its bedrock spillway (Pells et al., 2015). At the Oroville Dam in California (2017), breach of the emergency spillway produced a bedrock canyon tens of meters deep and hundreds of meters long through plucking in a single event (Nalin & Kotulla, 2019; Schmidt et al., 2017).

These dam analogs are significant because they document erosion at velocity scales (15–25 m/s) that overlap with or fall below the lower bound of this modeled fjord velocities. In open spillway geometries — where turbulent energy can disperse laterally — velocities of 25 m/s produce scour depths of tens of meters per event. In confined fjord channels, where turbulent energy is trapped between walls and concentrated at the bed, erosive capacity per unit velocity should be higher.

Bedrock Erosion: Transient Pulse vs. Sustained Flow

The megaflood analogs involved sustained flows of hours to weeks. A tsunami provides transient high-velocity flow — perhaps 10–30 minutes of extreme velocities at any given point, followed by backflow of comparable duration.

This disanalogy is less significant than it initially appears, for four reasons:

First, plucking is threshold-dependent, not cumulative. Once hydraulic forces exceed the resistance of a jointed block, the block is removed in seconds. The dam spillway literature confirms this: at Wivenhoe, 1,200-tonne boulders were removed during a single flood event at moderate velocities (Stratford et al., 2013). The rate-limiting step is not the time required to remove an individual block, but the rate at which flow accesses successive layers of blocks.

Second, dynamic pressure propagation into joint networks operates ahead of the erosion front (Bollaert & Schleiss, 2005; Pan et al., 2014). High-velocity flow penetrates joints meters ahead of the exposed surface, pre-loosening blocks that are removed when the erosion front reaches them. This means the effective erosion depth per pulse extends beyond the immediately exposed surface layer.

Third, a bolide swarm — multiple fragments impacting over hours — generates multiple tsunami arrivals, each producing a high-velocity pulse. Ten to twenty pulses over several hours, with each backflow phase also exceeding erosion thresholds, produces an event sequence that compounds erosion progressively. The cumulative progressive failure mechanism documented by Pan et al. (2014) — in which repeated pressure cycles weaken joint surfaces even below the single-cycle plucking threshold — means that later pulses in the sequence are more erosively efficient than earlier pulses.

Fourth, progressive channel deepening during the event creates a positive feedback: as the channel deepens, hydraulic focusing increases, concentrating more energy per unit surface area on the bed. Each successive pulse in the sequence acts on a deeper, narrower channel geometry. The Kariba Dam plunge pool demonstrates this feedback at engineering scales: cumulative scour reached ~90 m below the tailwater level as progressive deepening concentrated spillway energy into an increasingly confined scour hole (Bollaert, 2004; Noret et al., 2012).

Quantitative Estimate

The dam spillway data provide a basis for bounding per-pulse erosion depth. At Wivenhoe (velocities <25 m/s, open geometry), scour reached tens of meters. At Kariba (repeated operation at comparable velocities), cumulative scour reached ~90 m. These modeled velocities in confined geometry represent shear stresses 1–10× greater than these dam analogs. A conservative per-pulse estimate of 10–50 m of bedrock removal in densely jointed gneiss is consistent with the dam engineering data, with higher values for later pulses in the sequence due to progressive joint weakening and channel deepening feedback.

For an event sequence of 10–20 pulses (including backflow phases) over several hours, total deepening of 200–1,000 m is within the range indicated by the physics. Starting from tectonic fracture valleys of 100–300 m structural relief — typical of the Caledonian orogen in western Norway — total post-event depths of 300–1,300 m are achievable. Sognefjord (~1,500 m) represents the upper bound of this range and may reflect locally favorable conditions including unusually dense jointing, optimal channel geometry for hydraulic focusing, or a pre-existing valley somewhat deeper than average.

This estimate is a first-order bounding calculation, not a computational model. The actual erosion achievable depends on variables including joint spacing and orientation in Norwegian gneiss, the specific bore dynamics in fjord geometry, debris transport capacity of the flow, and the partitioning of energy between ice removal and bedrock erosion. Computational modeling of bore propagation and plucking in fjord geometry would substantially refine these estimates.

Biological Evidence Consistent with the Model

Deep-Water Emergence in Norwegian Fjords

Norwegian fjords contain an anomalous concentration of deep-sea organisms at shallow depths — a phenomenon termed deep-water emergence (DWE). Norway has the most extreme DWE documented globally (Häussermann et al., 2021). The cold-water coral *Lophelia pertusa*, whose normal depth range is 200–1,000 m, occurs as shallow as 39 m in the Trondheimsfjord — the shallowest known occurrence worldwide (Fosså et al., 2002). Norway's estimated 2,000 km^2 of *Lophelia* reef represents the highest known density of this species anywhere in its range. The Hardangerfjord alone hosts six distinct deep-water biotopes between 100 and 400 m: *Lophelia* reef, sponge garden, seapen communities, soft-bottom coral garden, hard-bottom coral garden, and cerianthid stands (Buhl-Mortensen & Mortensen, 2014). The bamboo coral *Isidella lofotensis* is "rarely found outside Norwegian waters" and occurs in dense stands (up to 167 colonies per 100 m^2) in fjord basins. Live deep-sea foraminifera have been documented in the 850 m deep Hardangerfjord, with diversity "comparable both to that of the open deep sea and that of reported macrofauna from the same sites" — described as "a unique environment with elements of deep-sea faunas in a land-locked setting" (Austin & Howe, 2010). These deep-water populations are self-sustaining breeding communities, not transient strays (Häussermann et al., 2021).

The standard explanation attributes DWE to fjord geometry: deep basins behind shallow sills create environmental conditions (cold, dark, saline) suitable for deep-water species, whose larvae arrive via inflowing ocean water. This accounts for the maintenance of existing populations but raises questions about initial colonization. Gene flow between Norwegian *Lophelia* populations is "sporadic" and recolonization through larval dispersal is "likely to take a long time" (Le Goff-Vitry & Rogers, 2005). Yet coral reef ecosystems "widely established themselves across the entire 3,000 km Scandinavian shelf prior to ~10 ka BP" (López Correa et al., 2012), within approximately 1,600 years of ice retreat and the establishment of Atlantic water circulation — with accretion rates up to 614 cm/ka, two to three times faster than comparable reefs elsewhere. Under this model, the proposed event at ~12,800 BP precedes coral establishment by approximately 2,800 years; however, the intervening Younger Dryas cold period (12,800–11,700 BP) involved ice readvance over the outer fjord regions and conditions unfavorable for coral growth (López Correa et al., 2012). The relevant timescale is therefore from the restoration of suitable conditions (~11,700 BP) to established reefs (~10,000 BP) — approximately 1,700 years, during which any seeded material would have persisted as dead substrate or dormant propagules in anoxic basin sediments. This rapid colonization despite restricted larval transport is difficult to reconcile with the gradual model. Moreover, while DWE occurs in fjords globally (Chile, New Zealand, Canada, Alaska), nowhere is it as extreme as in Norway (Häussermann et al., 2021). If all fjords create suitable conditions through geometry, the anomalous intensity of Norwegian DWE requires additional explanation.

Tsunami Biological Transport: The Documented Mechanism

Tsunamis demonstrably transport marine biological material over large distances. Deep-water foraminifera are a primary diagnostic indicator of tsunami deposits, with species from "below the storm wave base" serving to distinguish tsunami from storm deposits (Mamo et al., 2009; Pilarczyk et al., 2020). The 2004 Indian Ocean tsunami deposited benthic foraminifera from shelf depths of 30–80 m onshore (Bahlburg & Weiss, 2007; Hawkes et al., 2007). Tsunami deposits characteristically contain chaotic assemblages mixing organisms from deep, shallow, brackish, and freshwater environments — a diagnostic signature not produced by other processes.

The Storegga submarine slide (~8,150 BP) provides a direct analog on the Norwegian coast. Bondevik et al. (1997) documented Storegga tsunami deposits in more than 40 coastal lakes along western Norway, containing marine diatoms, shell fragments, sea urchin fragments, and marine foraminifera — material demonstrably transported from the sea into freshwater basins. Recent work has substantially expanded the known footprint of this event. Bondevik et al. (2024) showed that the Storegga tsunami reworked sediments across "large parts of the sea floor in the Norwegian and North Seas," with flow velocities of 2–5 m/s on the shelf offshore western Norway. Foraminifera in the reworked layer were radiocarbon dated to 11,000 years BP — the tsunami excavated and redeposited material thousands of years older than itself. Most remarkably, Dobrovolskas et al. (2025) identified Storegga tsunami deposits 1,250 km from the source in the NW Barents Sea (75°N), containing foraminifera dated 12,000–27,000 years BP along with terrestrial plant sedaDNA and terrestrial steroids.

The Colonization Puzzle Across the Western Fjord Zone

The deep-water biological anomaly is not confined to a single fjord. Across the entire western Norwegian fjord zone, researchers have documented deep-sea organisms inhabiting fjord basins that are physically isolated from the open deep ocean by shallow sills — and in each case, the mechanism of initial colonization is identified as unresolved.

Sognefjord — described as "the inland deep sea" by Buhl-Mortensen et al. (2020) — contains an assemblage of continental-slope species behind a sill approximately 100 m deep. The deep-sea soft coral *Anthomastus grandiflorus*, whose documented depth distribution is 457–1,760 m on the continental slope (Molodtsova et al., 2008), occurs at 540 m inside the fjord. The deep-sea holothurian *Bathyplotes natans* and the burrowing decapods *Munida sarsi* and *Munida tenuimana* dominate the central fjord basin (Bernd, 1993; Buhl-Mortensen et al., 2020). The deep-sea pennatulaceans *Kophobelemnon stelliferum* and *Virgularia mirabilis* occur below 630 m and 500 m respectively. Buhl-Mortensen et al. (2020) note that "the isolation of fjord basins behind one or more sills can lead to mass occurrence of species that by chance have been introduced and been able to establish dense populations" — acknowledging that the colonization mechanism is unknown and attributing it to chance.

In Hardangerfjord (860 m, sill at 150–200 m), the deep-sea crinoid *Conocrinus lofotensis* — a quintessentially bathyal organism — inhabits the inner basin behind the moraine sill (Buhl-Mortensen & Buhl-Mortensen, 2014). Austin and Howe (2010) documented live deep-sea benthic foraminifera in the 850 m basins, with species diversity "comparable both to that of the open deep sea and that of reported macrofauna from the same sites." They described the fjord as "a unique environment with elements of deep-sea faunas in a land-locked setting" and explicitly questioned the means of colonization given "the isolation of the deepest basin from the open deep sea both by the relatively shallow continental shelf and the presence of sills within the fjord system."

A pattern emerges from comparing these fjords: the depth of the fjord basin predicts which deep-sea species are present. Sognefjord (1,308 m) contains continental-slope species at 500–630 m depth that are absent from Hardangerfjord (860 m), despite both fjords having sills of similar depth (100–200 m) and receiving larvae from the same coastal water mass (Buhl-Mortensen et al., 2020). If colonization occurred through gradual larval transport over the sill, both fjords should host similar communities — larvae entering over a 100–200 m sill cannot selectively deliver 500 m species to one fjord and 200 m species to another. The fact that basin depth, not sill depth, predicts community composition indicates that the organisms were deposited at depth rather than filtered through the sill from above.

The genetic evidence reinforces this interpretation across multiple taxa. Fjordic *Lophelia pertusa* populations are "highly differentiated genetically," with individual subpopulations such as the Osterfjord colony showing "very low genetic diversity" (Le Goff-Vitry & Rogers, 2005). The glacier lanternfish *Benthosema glaciale* — a deep-water mesopelagic fish — shows "contrasting patterns in fjords and the ocean" in distribution, growth, and population genetics, with fjord populations genetically differentiated from offshore populations (Kristoffersen & Salvanes, 1998). In both cases — sessile coral and mobile fish — fjord populations are genetically isolated from their oceanic source, consistent with founder-effect colonization from a single introduction event rather than ongoing gene flow.

Under the catastrophic model, these observations resolve into a single explanation. A mega-tsunami overtopping the sill at the fjord mouth carries deep-ocean water — and the organisms suspended in it — directly into the fjord basins. The depth of water transported determines which organisms arrive: a surge reaching 500+ m depth offshore would carry continental-slope fauna into Sognefjord's deep basins while a shallower surge would deposit only upper-slope species in Hardangerfjord. The low genetic diversity and high population differentiation of fjordic populations are the signature of founder-effect colonization — each fjord received a small, random sample of the offshore population in a single event, then reproduced in isolation. The consistency of this pattern across the entire 59–63°N zone is predicted by a coast-wide catastrophic event and is difficult to reconcile with independent, gradual colonization of each fjord by the same unidentified mechanism.

Anthomastus grandiflorus — the species found at 540 m inside Sognefjord — has been documented at the Charlie-Gibbs Fracture Zone on the Mid-Atlantic Ridge (Molodtsova et al., 2008), the same oceanic feature this paper identifies as the trans-Atlantic energy pathway for the Younger Dryas mega-tsunami. The organism and the proposed wave share a common source geography.

Preservation Over Millennia

The Storegga deposits also demonstrate that biological material transported by tsunami can survive thousands of years in fjord and coastal sediments. Bondevik et al. (2006) recovered green-colored moss stems from Storegga deposits that still contained intact chlorophyll after 8,000 years — preserved by rapid burial in shell-bearing sand that provided high pH, subsequently sealed by marine silt preventing oxygen and light exposure. More broadly, Norwegian fjord sediments preserve organic matter at rates twice the global ocean average per unit area, burying approximately 11% of annual marine carbon globally despite the small total area of fjords (Smith et al., 2015). Anoxic conditions behind fjord sills dramatically enhance preservation of organic compounds that would degrade rapidly in oxygenated settings (Henrichs, 1992).

From Møre og Romsdal to a General Mechanism: Bolide Impact as the Origin of Fjords

The preceding sections developed a specific physical model for the catastrophic formation of Møre og Romsdal fjords by Younger Dryas bolide-generated mega-tsunami at ~12,800 BP. During the course of this investigation, a broader pattern emerged: Norwegian fjords do not share a single orientation, and the morphological character of each orientation group correlates with the age of confirmed impact structures in the water body each group faces. This section presents that pattern and its implications for fjord formation as a general mechanism.

Fjord Orientation Groups in Norway

Norwegian fjords fall into four distinct orientation groups. The western fjords of Møre og Romsdal and Vestland — including Sognefjord, Hardangerfjord, Romsdalsfjord, Storfjord, and Nordfjord — trend roughly east-west, opening directly onto the Atlantic Ocean. The northern fjords of Nordland and Troms — including Ofotfjord, Lyngenfjord, and the smaller fjords threading through the Lofoten–Vesterålen archipelago — trend northwest-southeast, facing the Norwegian Sea. The Finnmark fjords — Porsangerfjord, Tanafjord, Laksefjord opening north, and Varangerfjord opening east — face the Barents Sea (MAREANO, 2014). The Oslofjord and adjacent Skagerrak-facing coast are oriented south toward the North Sea basin.

Depth-Exposure Gradient

The depth of Norwegian fjords correlates systematically with the degree of open-ocean exposure. The western Atlantic-facing fjords are the deepest: Sognefjord reaches 1,308 m water depth with bedrock at approximately 1,500 m below sea level (Nesje et al., 1992; Nesje & Whillans, 1994), Hardangerfjord reaches approximately 860 m (Hjelstuen et al., 2009), and Romsdalsfjord approximately 800 m. The fjords facing the Norwegian Sea are intermediate: Trondheimsfjord reaches approximately 617 m, Ofotfjord 553 m (Norwegian Hydrographic Service). The Finnmark fjords facing the Barents Sea are the shallowest and morphologically least fjord-like: Altafjord reaches 488 m but is 7 km wide, and Porsangerfjord — Norway's fourth-longest at 123 km — has entrance depths of approximately 250–300 m over a relatively flat bottom (Stramska et al., 2018). The MAREANO seabed mapping program describes the Finnmark coast as "dominated by wide fjords which are almost like ocean areas" that, "in contrast to the fjords in west Norway, often lack a threshold at their entrances" (MAREANO, 2014). The Oslofjord, facing the sheltered Skagerrak, reaches only approximately 258 m and is classified as a rift valley rather than a glacially carved fjord (Larsen, 2008).

The standard glacial model attributes this gradient to ice dynamics: channeled outlet glaciers draining steep western terrain erode more deeply than the broad, uniform ice sheet that covered the low-relief Finnmark plateau (Syvitski et al., 1987). This is a legitimate factor. However, a critical lithological observation complicates this explanation. The bedrock along the Finnmark coast consists of "mainly old and hard metamorphic rocks such as gneiss, quartzite and schist" (MAREANO, 2014) — the same assemblage that characterizes the western Norwegian fjord region (Nesje & Whillans, 1994; Sigmond et al., 1984). If bedrock resistance were equal, as the lithological data indicate, the glacial model must attribute the entire depth difference to ice dynamics alone. Bernard et al. (2021, GRL) demonstrated that lithology controls fjord width, with harder rocks producing narrower fjords, but found that depth is controlled by other factors — leaving the depth-exposure correlation unexplained by lithology.

The Sognefjord Structural Anomaly

Sognefjord — the deepest fjord in Norway — presents a specific structural puzzle. Nesje and Whillans (1994) established that the fjord system follows zones of rock-structural weakness, but the main fjord trends east-west while the Caledonian fold structure in the region strikes northeast-southwest (Nicholson, 1963; Nesje & Whillans, 1994). The fjord's primary axis is not aligned with the dominant bedrock weakness. This contrasts with Hardangerfjord, which follows the Hardangerfjord Shear Zone (Fossen & Hurich, 2005), and Lyngenfjord, whose direction corresponds to the local fracture pattern (Randall, 1961). Sognefjord points directly west toward the open Atlantic, independent of the geological structure through which it cuts. The glacial model explains this by invoking the pre-glacial river valley that guided ice flow, but Nesje and Whillans (1994) reconstructed the pre-glacial valley as a gentle feature approximately 500 m above present sea level — not a deep channel that would strongly constrain ice flow direction.

The Sediment Volume Discrepancy

An independent observation bears on this question. Steer et al. (2012, Nature Geoscience) compared fjord erosion volumes in western Scandinavia with offshore sediment deposition during the Plio-Quaternary (0–2.8 Ma) and found that fjord erosion accounts for only 35–55% of the total deposited sediment volume. The remaining 45–65% was attributed to erosion of the low-relief plateau surfaces between fjords. Under the catastrophic model, a substantial fraction of the offshore sediment attributed to gradual plateau erosion may instead represent catastrophic excavation deposits — material ejected from fjords during the carving event and deposited on the shelf in a single stratigraphic horizon. This is consistent with the "extensive re-sedimentation" documented at the YD horizon in the Nordfjord system (Hjelstuen et al., 2009) and the unsorted, mixed-lithology deposits described by Holtedahl (1959) on the Norwegian shelf at the ~13,000 BP level.

Impact Structures in Every Facing Water Body

Each water body faced by a Norwegian fjord group contains a confirmed or candidate impact structure.

The Barents Sea — faced by the Finnmark fjords — contains the Mjølnir crater, a confirmed 40-km-diameter complex impact structure produced by a 1.5–2.0 km bolide at approximately 142 Ma (Gudlaugsson, 1993; Dypvik et al., 1996; Tsikalas et al., 1999). The impact occurred in a shallow epicontinental sea of 300–500 m depth, and the crater is preserved beneath 50–150 m of post-impact sediment (Dypvik et al., 2004).

The North Sea — toward which the Oslofjord and Skagerrak coast are oriented — contains the Silverpit crater, a 3-km impact structure surrounded by a 20-km ring of circular faults. Its impact origin was confirmed in 2025 through shocked mineral grains, high-resolution 3D seismic imaging, and numerical impact simulations, with biostratigraphic dating to the middle Eocene at 43–46 Ma (Nicholson et al., 2025, Nature Communications). Silverpit had been disputed for over two decades, with a 2009 Geological Society of London debate voting overwhelmingly against impact origin — a cautionary example of how marine impact structures can be dismissed without definitive petrographic evidence.

The Norwegian Sea — faced by the Nordland and Troms fjords — has no confirmed impact structure. However, Riis et al. (2005) identified several crater-like structures up to 20 km wide and 300 m deep in the Møre Basin, formed in Oligocene–Miocene ooze sediments. These are currently attributed to sediment evacuation triggered by mass flows, but they have not been tested for impact indicators. The Norwegian Sea deep basin has received substantially less seismic survey coverage than the petroleum-rich Barents Sea and North Sea shelves. Given that Silverpit required 23 years from discovery to confirmation, the absence of a confirmed impact structure in the Norwegian Sea reflects the state of investigation, not a geological constraint.

The open Atlantic — faced by the western Møre og Romsdal and Vestland fjords — is the pathway for the Younger Dryas bolide-generated mega-tsunami at ~12,800 BP, the subject of this paper's primary physical model.

Age-Morphology Correlation

The morphological character of each fjord group correlates with the age of the impact in its facing water body, and this correlation follows the direction predicted by the catastrophic model: older impacts correspond to more degraded features.

The western fjords, corresponding to the youngest proposed event (~12,800 BP), are the deepest (1,308 m), narrowest (Sognefjord averages <5 km width), steepest-walled, and possess the most pronounced sills — the morphological signature of a recently carved channel with minimal post-formation modification.

The Finnmark fjords, corresponding to the oldest confirmed impact (Mjølnir, 142 Ma), are the most morphologically degraded. They are broad, shallow relative to their length, lack steep walls, and frequently lack sill thresholds (MAREANO, 2014). The scientific community does not universally classify Finnmark inlets as fjords in the strict geomorphological sense (Syvitski et al., 1987). This is consistent with 142 million years of sedimentary infill, isostatic adjustment, tectonic modification, and 20+ Quaternary glacial cycles progressively degrading steep-walled channels into broad, shallow inlets.

The Nordland and Troms fjords are morphologically intermediate — deep and steep-walled with alpine character, but generally less extreme than the western group. Their facing water body (the Norwegian Sea) has no dated impact, precluding placement on the age-morphology axis, but the intermediate morphology is consistent with an intermediate-age source event.

Global Fjord Distribution and Impact History

The pattern observed in Norway — fjords facing water bodies with confirmed impact structures, depth correlating with ocean exposure — is consistent with the global distribution of fjords. Every major fjord coastline faces a vast open ocean basin, and the detection record for oceanic impacts is severely incomplete: of approximately 190 confirmed impact structures in the Earth Impact Database, fewer than 33 are submarine, despite at least 70% of all bolide impacts striking ocean surfaces (Nicholson et al., 2024). Oceanic crust older than approximately 200 Ma has been recycled through subduction, permanently erasing earlier impact evidence. Deep water can absorb impact energy without producing a seafloor crater, as demonstrated by the Eltanin impact (Ward & Asphaug, 2002). The absence of known oceanic impact craters adjacent to a fjord coastline reflects a detection limitation, not a geological constraint.

Examination of the terrestrial and near-shore record reveals that every major fjord region has documented Quaternary-age impact events, impact proxy evidence, or candidate structures in its vicinity.

Chile's Pacific fjord coast faces the ocean basin where the Eltanin asteroid impact — a confirmed craterless deep-ocean event at approximately 2.51 Ma — generated mega-tsunamis exceeding 200 m at proximal coastlines and reaching tens of meters at distances exceeding 5,000 km (Ward & Asphaug, 2002). Goff et al. (2012) identified possible Eltanin megatsunami deposits in Antarctica, Chile, and New Zealand, and noted that the event's timing coincides with the Pliocene-Pleistocene boundary — the onset of the glacial cycles conventionally invoked to explain fjord formation. The Bajada del Diablo crater field in Argentine Patagonia (~42°45′S), at the latitude of the northern Chilean fjord zone, contains nearly 200 impact structures dated between 0.13 and 0.78 Ma (Acevedo et al., 2011). At the Pilauco site in southern Chile (~40°S), peak Younger Dryas boundary concentrations of platinum, gold, high-temperature spherules, and native iron particles indicate cometary impact at approximately 12,800 BP (Pino et al., 2019).

New Zealand's Fiordland faces the Southern Ocean where the Eltanin impact modeling indicates 60 m waves reached the New Zealand coast (Ward & Asphaug, 2002). Across the Tasman Sea, the Darwin Crater in western Tasmania — a confirmed Quaternary impact structure dated to 816 ± 7 ka — distributed impact glass across 400 km^2 (Ford, 1972; Lo et al., 2002). Tasmania also lies within the Australasian tektite strewn field, the largest on Earth, produced by an impact at approximately 770–785 ka whose source crater was identified only recently beneath lavas in southern Laos (Sieh et al., 2019). Records exist of a submarine bolide impact near Stewart Island adjacent to Fiordland (Nathan, 2009). Fiordland itself is essentially unstudied for paleotsunamis.

Alaska's fjord coast has the Sithylemenkat Lake candidate impact crater in central Alaska, identified as "formed near the end of the Wisconsinan glaciation" based on radial and concentric fracture patterns and anomalous nickel concentrations (Cannon, 1977; Davis, 1977). Two additional suspected impact structures exist in Alaska: Savonoski Crater in Katmai and a feature on Amak Island (Davis, 1977).

Scotland's sea lochs — fjords by the geomorphological definition — occur almost exclusively on the Atlantic-facing west coast, despite glacial outlet activity on both coasts. This directional selectivity is predicted by the catastrophic model (Atlantic-facing = maximum wave exposure) but not naturally explained by glacial processes alone.

The Eltanin precedent merits particular emphasis. This confirmed deep-ocean asteroid impact at 2.51 Ma left no crater, is detectable only through indirect evidence in ocean cores, and delivered continental-scale wave energy precisely to the ocean-facing coastlines where fjords are concentrated — Chile, New Zealand, and Antarctica. Its timing at the Pliocene-Pleistocene boundary raises the question of whether the onset of "glacial" fjord formation was in fact triggered by impact-generated mega-tsunami excavation of coastal valleys, with subsequent glacial cycles modifying rather than creating the resulting channels.

Implications

If bolide impacts are a general mechanism for fjord formation, the conventional model requires revision not in its description of glacial modification — which is real and well-documented — but in its attribution of primary excavation to glacial processes. Under the multi-event model, the sequence is: (1) tectonic fracture valleys form the initial template; (2) oceanic bolide impacts generate mega-tsunamis that catastrophically excavate these valleys into fjords; (3) subsequent glacial cycles deposit ice in the pre-existing channels, producing glacial striations, moraines, and other modification features; (4) deglaciation floods the channels with seawater. Different fjords may have been carved by different impact events at different times, explaining why fjord morphology varies regionally in ways that correlate with impact age rather than with uniform Quaternary glacial history.

This multi-event framework is preliminary beyond the Møre og Romsdal–Boston Basin connection. The Mjølnir and Silverpit impacts are confirmed, but their role in excavating Finnmark and south Norwegian features has not been modeled. The Norwegian Sea gap remains. The age-morphology correlation could reflect confounding variables including differential topographic relief, ice sheet configuration, and tectonic history. The pattern is presented as an observation that warrants systematic investigation and that, if confirmed, would transform the understanding of fjord formation from a glacial erosion problem to an impact geomorphology problem.

Discussion

The catastrophic hydraulic model addresses several features of Norwegian fjords that the glacial model accommodates with difficulty: the directional concentration on Atlantic-facing coasts; the depth-exposure gradient across fjord groups despite uniform gneissic bedrock; the structural independence of Sognefjord's axis from the Caledonian fold pattern; the sediment volume discrepancy identified by Steer et al. (2012); and the biological colonization puzzle in which deep-sea species inhabit fjord basins isolated by shallow sills through an unidentified mechanism. It does so using well-documented physical mechanisms — hydraulic focusing and threshold-dependent plucking — applied at energy scales consistent with independent evidence for the Younger Dryas impact.

Parallel Anomalies at Source and Receiving End

Both ends of the proposed trans-Atlantic propagation path exhibit the same pattern: YD-age geological features that the glacial model cannot explain, documented by each region's own geological experts independently of this paper.

At the proposed source site, the Boston Basin presents a suite of anomalies that Gjøvik (2026) argues are diagnostic of bolide impact: fault-controlled kaolinization to depths exceeding 300 feet — a hydrothermal signature that Kaye (1967) described as "conjectural" and could not explain; rock recrystallization at 175–250°C with no documented heat source for over 400 million years (Anderson, 2008); fresh olivine in the Main Drainage Tunnel alongside partially altered chlorite and magnetite (Rahm, 1962) — olivine is among the least stable rock-forming minerals at surface conditions and cannot persist for 400 Ma without recent emplacement; a chondritic mineral assemblage including albitic feldspar (Ab_{84}), ilmenite in three alteration stages, chromite with Cr at 1,763 ppm and Ni at 1,040 ppm (Thompson, 2020; Ross, 1990); and platinum group elements at five times crustal background in harbor sediments that did not decrease with cessation of anthropogenic sludge input (Tuit, 2000). Boston's own geological community dates the mass debris event (Lexington Substage) to ~12,900 BP but acknowledges that the readvance "only extended into valleys north and west of Boston" (Barosh, 2011) — the features in Boston proper formed without glaciers present. The basin has been described as "probably one of the last to be mapped satisfactorily" (Kaye, 1984) with surficial deposits constituting "probably the most complex in the country" (Barosh, 2011). No consensus explanation exists for the geology of Boston despite over a century of study.

At the proposed receiving end, the Sunnmøre-Nordfjord region of Møre og Romsdal — the most Atlantic-exposed section of the Norwegian coast — shows significantly less YD readvance than regions both to the north and south (Mangerud et al., 1979), despite receiving comparable or equivalent precipitation (Larsen et al., 1984). No specific mechanism has been proposed in the literature to account for this regional anomaly.

In both locations, the explanatory vacuum predates and is independent of the catastrophic hypothesis. The model connects them: energy source at one end, energy delivery at the other, with the Charlie-Gibbs Fracture Zone as the preferential trans-Atlantic pathway between them.

The Drumlin Swarm and Wave Generation

The Boston Basin drumlin field provides a quantitative constraint on the swarm that previous YDIH literature lacked. La Forge (1932) documented over 200 drumlins in eastern Massachusetts. The field extends at least 16 km seaward into Massachusetts Bay (Rosen 2004), and Rosen notes that "what we see on land is a small fraction of the drumlin field that once existed" — most are now submerged. Under the Gjøvik (2026) interpretation, each drumlin is an ejecta mound from an individual fragment impact, giving a minimum fragment count of approximately 200 across a ~50 km zone.

The 34 drumlin islands in Boston Harbor (NPS, 2017) indicate that 30–50 fragments impacted zones where the bedrock is at or near present sea level — either open shallow marine, thin ice over water, or the ice-water margin of the basin's offshore extension into Massachusetts Bay. These marine-impacting fragments generate waves through direct water displacement, the most efficient conversion mechanism, with shallow-water impacts producing higher coupling efficiency than deep-ocean impacts (Ward & Asphaug, 2000).

Three independent wave generation mechanisms operate simultaneously in this scenario. The marine-impacting fraction (30–50 fragments in shallow water) generates tsunamis through direct displacement — even at conservative 1% coupling efficiency, 40 fragments averaging 50 m diameter produce wave energy comparable to the 2004 Indian Ocean tsunami (~1.1×10^{17} J; USGS). At the higher coupling efficiencies expected for shallow-water impacts, this fraction alone generates 5–10× the Indian Ocean tsunami wave energy. The ice-sheet-impacting fraction (~150–170 fragments through 2–3 km of ice) vaporizes and melts approximately 8–9 km^3 of ice, releasing this water effectively instantaneously from 2–3 km elevation onto the continental shelf — a catastrophic meltwater cascade that generates its own wave train independent of the impact waves. And the atmospheric blast from ~200 simultaneous vaporization events at total swarm energy of approximately 5.7×10^{19} J (~68× Krakatoa) generates waves of 8–20 m at 5,500 km distance through atmospheric coupling alone, using the impulse-dependent scaling bounded in Section 4.2.

A critical geometric factor distinguishes this scenario from the point-source models used for calibration. A distributed swarm across ~50 km generates a quasi-planar wavefront that attenuates as $1/\sqrt{r}$ rather than the $1/r$ decay of a point-source radial wave. At trans-Atlantic distances (~5,500 km), this geometric advantage preserves substantially more wave amplitude than a point source of equivalent total energy. The Ward (2003) calibration — based on a single deep-ocean impact — is therefore conservative for a distributed shelf-margin swarm.

Combined arrival estimates at the Norwegian outer coast, accounting for all three mechanisms and the distributed-source geometry: 15–40 m before fjord amplification. After Green's Law amplification of ~4.5× in typical MOR geometry: 70–180 m inner fjord wave heights, corresponding to flow velocities of 30–85 m/s. These velocities exceed the catastrophic plucking threshold (5–10 m/s) by factors of 3–17× and overlap with or exceed the dam spillway velocities (15–25 m/s) that produced 90 m of scour at Kariba and "almost unbelievable" erosion at Wivenhoe (Bollaert, 2004; Stratford et al., 2013).

The North Atlantic Drumlin Corridor

The three partially submerged drumlin fields documented worldwide are Boston Harbor, Clew Bay on the Atlantic-facing west coast of Ireland, and Fláajökull in southeast Iceland (NASA Earth Observatory, 2025). All three are in the North Atlantic. Under the glacial model, this geographic clustering is coincidental. Under the impact model, it is the debris corridor.

Clew Bay lies directly in the CGFZ propagation path from Boston to Norway — precisely where Ward (2003) modeled 20–23 m wave arrival at Ireland. If the Boston Basin drumlins are impact ejecta rather than glacial features, then the Clew Bay drumlins warrant reexamination under the same framework. Their Atlantic-facing position, their partially submerged state, and their morphological similarity to the Boston features are consistent with either secondary impact debris from the swarm's distal fragments or tsunami-deposited material from the Boston event. Iceland's Fláajökull field — also Atlantic-facing — presents the same question. Three anomalous drumlin fields, all on the Atlantic, all on the wave propagation path between the proposed impact site and the Norwegian coast, is a pattern that the glacial model does not predict and the impact model does.

The Impact Orientation

The observation that Norwegian fjord groups face water bodies containing confirmed impact structures, with morphological character correlating to impact age, has implications that extend beyond Norway.

The Finnmark-Mjølnir connection is the most immediately accessible. The Mjølnir crater is well-characterized and the Finnmark shelf has been surveyed extensively through petroleum exploration. If Finnmark fjord sediments at or near their basal contact contain signatures consistent with impact-generated tsunami deposition — chaotic assemblages, mixed-depth organisms, poorly sorted lithologies — this would constitute independent evidence for the catastrophic model applied to a confirmed impact event. The challenge is temporal: 142 million years of subsequent geological processes may have overprinted any original depositional signature.

The Silverpit connection opens a different line of inquiry. Silverpit's confirmation as an impact structure in 2025 (Nicholson et al., 2025) means its potential geomorphological consequences have not been evaluated. South-facing features along the Skagerrak coast could be examined for consistency with a 43–46 Ma impact-tsunami event. Eocene sedimentary records are well-preserved in the North Sea basin.

The Norwegian Sea represents the model's most significant gap. The Riis et al. (2005) crater-like structures in the Møre Basin — up to 20 km wide and 300 m deep — have been attributed to ooze evacuation but have not been examined for impact indicators. The Silverpit precedent is directly relevant: that structure was dismissed for over two decades before shocked minerals were identified in drill cuttings from a 1985 well (Nicholson et al., 2025). Examining the Riis et al. structures for shocked quartz or feldspar in existing core material would be straightforward and could either close or confirm the Norwegian Sea gap..

Biological Evidence

The cross-fjord colonization pattern — in which basin depth rather than sill depth predicts which deep-sea species are present — constitutes an independent line of evidence. Sognefjord (1,308 m) contains continental-slope species at 500–630 m depth that are absent from Hardangerfjord (860 m), despite both fjords having sills of similar depth (100–200 m) and receiving larvae from the same coastal water mass (Buhl-Mortensen et al., 2020). Under the catastrophic model, deeper tsunami surges deposit deeper-water organisms — producing a depth-fauna correlation that gradual larval transport over the sill cannot explain. A systematic comparison of deep-water fauna across Norwegian fjords of varying basin depth but similar sill depth has not been conducted but would be informative.

The founder-effect genetic signature in fjordic Lophelia (Le Goff-Vitry & Rogers, 2005) and Benthosema glaciale (Kristoffersen & Salvanes, 1998) populations reinforces this interpretation. If catastrophic transport seeded fjord populations in a single event, molecular clock analysis of divergence times between fjordic and oceanic populations should converge on a common date. If that date is consistent with ~12,800 BP for western fjords, the biological and geological evidence would be independently convergent. Existing genetic datasets may already contain this information — the analysis requires reframing the question from "how isolated are fjord populations" to "when did they diverge from their oceanic source.

Critical Uncertainties

The energy partitioning between ice-sheet-impacting and marine-impacting swarm fragments remains the model's most significant quantitative uncertainty. The drumlin-based swarm estimate partially resolves this by establishing that a substantial fraction of fragments impacted the marine or near-marine portion of the basin, but the wave generation efficiency of ice-sheet impacts has never been computationally modeled. Until this calculation is performed, wave height and flow velocity estimates remain conditional.

The transient-versus-sustained flow question represents a second uncertainty. While threshold-dependent plucking mechanics and the dam spillway literature support substantial erosion from transient high-velocity pulses, the direct comparison with sustained megaflood analogs is imperfect. Computational modeling of bore propagation and plucking in realistic fjord geometry would constrain the per-pulse erosion estimate.

The multiple-events question for the western Norwegian fjords is a third uncertainty. The Atlantic has experienced multiple impacts over geological time, and without direct dating of the bedrock erosion surface beneath fjord sediments, we cannot determine whether all western fjords were carved at ~12,800 BP or whether some record earlier events. Cosmogenic exposure dating of submerged fjord walls — technically challenging but not impossible — would resolve this definitively.

Conclusion

The physical model developed in this paper demonstrates that a trans-Atlantic bolide-generated mega-tsunami at

~12,800 BP could have carved the Møre og Romsdal fjords from pre-existing tectonic fracture valleys. The Boston Basin drumlin field constrains the swarm to approximately 200 fragments across a ~50 km zone, with 30–50 impacting shallow marine environment directly. Three independent wave generation mechanisms — direct marine displacement, catastrophic meltwater release from vaporized glacier, and atmospheric blast coupling — operate simultaneously, with the distributed swarm geometry preserving wave amplitude across trans-Atlantic distance more efficiently than a point source. Energy analysis yields inner-fjord flow velocities of 30–85 m/s, far exceeding the 5–10 m/s threshold for catastrophic bedrock plucking. Dam spillway engineering data from Kariba, Wivenhoe, Copeton, and Oroville confirm that erosion at these velocity scales produces results exceeding engineering predictions, and support per-pulse erosion estimates of 10–50 m in jointed metamorphic rock. Progressive joint weakening through repeated pressure cycles (Pan et al., 2014; Bollaert & Schleiss, 2005) and channel-deepening feedback compound erosion across a multi-pulse event, yielding total excavation consistent with observed fjord depths.

The model resolves anomalies that the standard glacial model does not: the directional concentration of deep fjords on Atlantic-exposed coasts; the structural independence of Sognefjord's axis from the bedrock fold pattern (Nesje & Whillans, 1994); the sediment volume discrepancy in which glacial fjord erosion accounts for only 35–55% of offshore deposition (Steer et al., 2012); the anomalously reduced YD readvance in the most Atlantic-exposed coastal section despite equivalent precipitation (Mangerud et al., 1979; Larsen et al., 1984); and massive undifferentiated resedimentation at the ~12,800 BP horizon that has never been examined using tsunami deposit diagnostic criteria (Hjelstuen et al., 2009; Holtedahl, 1959). Biological evidence — globally anomalous deep-water emergence (Häussermann et al., 2021), the colonization puzzle in which basin depth rather than sill depth predicts which deep-sea species are present (Buhl-Mortensen et al., 2020; Austin & Howe, 2010), founder-effect genetics across multiple taxa (Le Goff-Vitry & Rogers, 2005; Kristoffersen & Salvanes, 1998), and tsunami biological transport preserved for 8,000+ years on this coast (Bondevik et al., 2024; Dobrovolskas et al., 2025) — is consistent with catastrophic deep-marine deposition in fjord basins. The deep-sea soft coral Anthomastus grandiflorus, found at 540 m inside Sognefjord behind a 100 m sill, has been documented at the Charlie-Gibbs Fracture Zone (Molodtsova et al., 2008) — the same oceanic feature this paper identifies as the trans-Atlantic wave energy pathway.

The pattern extends beyond Møre og Romsdal. Norwegian fjord groups have distinct orientations, and every water body they face contains a confirmed or candidate impact structure — the Barents Sea (Mjølnir, 142 Ma), the North Sea (Silverpit, confirmed 2025 at 43–46 Ma), the Norwegian Sea (Riis et al. crater-like structures, untested for impact origin), and the Atlantic. The morphological character of each group correlates with impact age despite uniform gneissic bedrock across regions (MAREANO, 2014; Sigmond et al., 1984). The three partially submerged drumlin fields documented worldwide — Boston Harbor, Clew Bay on the Atlantic-facing west coast of Ireland, and Fláajökull in Iceland — all lie on the North Atlantic propagation path, with Clew Bay positioned precisely where Ward (2003) modeled 20–23 m wave arrival. Globally, every major fjord coastline faces an ocean basin with documented Quaternary-age impact events, from the Eltanin impact delivering 200+ m waves to Chilean and New Zealand fjord coasts (Ward & Asphaug, 2002) to the Sithylemenkat Lake candidate crater in Alaska at the terminal Wisconsinan (Cannon, 1977). The Eltanin impact at 2.51 Ma — confirmed, craterless, and coincident with the Pliocene-Pleistocene boundary — raises the question of whether the onset of "glacial" fjord formation was itself triggered by impact-generated mega-tsunami excavation.

This paper presents the Møre og Romsdal case in detail. The Norwegian orientation-impact correlation and the global pattern are presented as observations that emerged from this research and that warrant independent investigation. This paper calls on researchers working at fjord coastlines worldwide — particularly in Chile, New Zealand, Alaska, and Scotland — to evaluate the bolide impact mechanism against the geological evidence at their sites. If Norwegian shelf sediments at the ~12,800 BP horizon are demonstrably consistent with glaciomarine deposition and inconsistent with catastrophic emplacement, the Younger Dryas component of the model is falsified. The archived core material required to evaluate this condition exists.

Acknowledgements

The author conceptualized the study, developed the theoretical framework, and conducted all analyses. This research received no external funding and was conducted using personally acquired resources. No new data was created for this article.

The author acknowledges Claude AI (Opus 4.5) for their assistance with manuscript preparation, data interpretation

discussions, and research development over the course of this investigation

The author discloses that her direct ancestors have been documented in and around Tingvoll, Møre og Romsdal, Norway, since at least the early 18th century. The author's personal interest in her ancestry and prior identification of the Boston Basin as a candidate YD-age impact site (Gjøvik, 2026) preceded and motivated the development of the trans-Atlantic propagation model presented here. The personal connection to both ends of the proposed event trajectory is disclosed as a matter of transparency; however, the physical arguments stand or fall on their own merits.

References

Aarseth, I. 1997. Western Norwegian fjord sediments: age, volume, stratigraphy, and role as temporary depository during glacial cycles. *Marine Geology* 143: 39–53.

Aarseth, I. & Mangerud, J. 1974. Younger Dryas end moraines between Hardangerfjorden and Sognefjorden, Western Norway. *Boreas* 3: 3–22.

Acevedo, R.D., et al. 2011. Meteorite Impact Craters in South America: A Brief Review. Revista Brasileira de *Geomorfologia* 12(3): 137–160.

Andersen, B.G. et al. 1995. Younger Dryas ice-marginal deposits in Norway. *Quaternary International* 28: 147–169.

Anderson, E. P. (2008). Petrographic and SEM-EDS Analysis of Aspidella bearing Siltstones and Slates of the Cambridge Argillite, Boston Bay Group, Massachusetts. *21st Keck Symposium Volume.* Keck Geology Consortium.

Aure, J. et al. 2007. Primary production enhancement by artificial upwelling in a western Norwegian fjord. *Marine Ecology Progress Series* 352: 39–52.

Austin, W.E.N. & Howe, J.A. 2010. Deep-sea benthic foraminifera in Hardangerfjord, Norway: An initial assessment. *Estuarine, Coastal and Shelf Science* 89: 162–168.

Baker, V.R. 1973. Paleohydrology of Lake Missoula flooding. *GSA Special Paper* 144.

Baker, V.R. 2009. The Channeled Scabland: A retrospective. *Ann. Rev. Earth Planet. Sci.* 37: 393–411.

Balter-Kennedy, A. et al. 2024. The LIS in southern New England. *Climate of the Past* 20: 2167–2190.

Barosh, P.J. 2011. *The Geology of Boston*. Geological Society of America Field Guide.

Baynes, E.R.C. et al. 2015. Erosion during extreme flood events dominates Holocene canyon evolution. *PNAS* 112(8): 2355–2360.

Bernard, T., et al. 2021. The impact of lithology on fjord morphology. *Geophysical Research Letters* 48: e2021GL093101.

BerndC. (1993). A television and photographic survey of megafaunal abundance in Central Sognefjorden, Western Norway.*Sarsia*781–8.

Billings, M.P. 1979. Boston Basin, Massachusetts. In *The Geology of Selected Areas in New England*, eds. Skehan and Murray. Kendall/Hunt.

Bollaert, E.F.R. 2004. A comprehensive model to evaluate scour formation in plunge pools. *International Journal on Hydropower & Dams.*

Bollaert, E.F.R. & Schleiss, A.J. 2005. Physically based model for evaluation of rock scour due to high-velocity jet impact. *Journal of Hydraulic Engineering* 131(3): 153–165.

Bondevik, S., Svendsen, J.I. & Mangerud, J. 1997. Tsunami sedimentary facies deposited by the Storegga tsunami in western Norway. *Sedimentology* 44: 1115–1131.

Bondevik, S. et al. 2005. The Storegga Slide tsunami. *Marine & Petroleum Geology* 22: 195–208.

Bondevik, S. et al. 2024. Contamination of 8.2 ka cold climate records by the Storegga tsunami in the Nordic Seas. *Nature Communications* 15: 2870.

Bosman, Adele & Basson, Gerrit. (2020). Physical model study of bedrock scour downstream of dams due to spillway plunging jets. *Journal of the South African Institution of Civil Engineers*. 62. 36-52.

Breivik, H.M. 2014. Palaeo-oceanographic development and human adaptive strategies in the Pleistocene–Holocene transition. *The Holocene* 24(11): 1478–1490.

Brunt, K.M., Okal, E.A. & MacAyeal, D.R. 2011. Antarctic ice-shelf calving triggered by the Honshu (Japan) earthquake and tsunami. *J. Glaciology* 57(205): 785–788.

Bugge, T., Befring, S., Belderson, R.H., Eidvin, T., Jansen, E., Kenyon, N.H., Holtedahl, H. & Sejrup, H.P. 1988. A giant three-stage submarine slide off Norway. *Geo-Marine Letters* 7: 191–198.

Buhl-Mortensen, L. & Mortensen, P.B. 2014. Diverse and vulnerable deep-water biotopes in the Hardangerfjord.

Marine Biology Research 10(3): 253–267.

Buhl-Mortensen, Lene & Buhl-Mortensen, Pål & Glenner, Henrik & Bamstedt, Ulf & Bakkeplass, Kjell. (2020). The inland deep sea—benthic biotopes in the Sognefjord., *Seafloor Geomorphology as Benthic Habitat* (pp.355-372).

Collier, J.S. et al. 2015. Streamlined islands and the English Channel megaflood. *Global & Planetary Change* 135: 190–206.

Cannon, P.J. 1977. Meteorite Impact Crater Discovered in Central Alaska with Landsat Imagery. *Science* 196(4296): 1322–1324.

Davis, N. Alaskan Meteorite Crater Discovery, June 27, 1977, UAF Geophysical Institute, https://www.gi.alaska.edu/alaska-science-forum/alaskan-meteorite-crater-discovery

Dahlgren, K.I.T., Vorren, T.O. & Laberg, J.S. 2002. Late Quaternary glacial development of the mid-Norwegian margin. *Marine and Petroleum Geology* 19: 1089–1113.

Dobrovolskas, D. et al. 2025. Expanding the footprint of the Storegga tsunami through new evidence from Arctic marine sediments. *Scientific Reports* 15: 23841.

Dypvik, H., et al. 2004. Impact breccia and ejecta from the Mjølnir crater in the Barents Sea. *Norwegian Journal of Geology* 84: 143–167.

Dypvik, H., et al. 1996. Mjølnir structure: An impact crater in the Barents Sea. *Geology* 24(9): 779–782.

Firestone, R.B. et al. 2007. Evidence for an extraterrestrial impact 12,900 years ago. *PNAS* 104(41): 16,016–16,021.

Fossen, H. & Hurich, C.A. 2005. The Hardangerfjord Shear Zone in SW Norway and the North Sea: A large-scale shear zone in the Baltic crust. *Journal of the Geological Society* 162: 675–687.

Fosså, J.H. et al. 2002. The deep-water coral *Lophelia pertusa* in Norwegian waters: distribution and fishery impacts. *Hydrobiologia* 471: 1–12.

Ford, R.J. 1972. A possible impact crater associated with Darwin Glass. Earth and Planetary Science Letters 16: 228–230.

Garcia-Castellanos, D. & O'Connor, J.E. 2018. Outburst floods provide erodability estimates. *Scientific Reports* 8: 10573.

Gisler, G. et al. 2018. Near and far-field hazards of asteroid impacts in oceans. *Acta Astronautica* 156: 308–314.

Glasser, N.F. et al. 2008. Structural, tectonic and glaciological controls on fjord landscapes. *Geomorphology* 105: 291–302.

Gjøvik, A.M. 2026. Identification of a New England Bolide Impact Site: A Geologic Reckoning with the Ground-Zero for the Younger Dryas Event. *The Journal of Decolonized Ecology and Evolution* 1(1). DOI: 10.5281/zenodo.19146482.

Goff, J., et al. 2012. The Eltanin asteroid impact: possible South Pacific palaeomegatsunami footprint and potential implications for the Pliocene–Pleistocene transition. Journal of Quaternary Science 27(7): 660–670.

Gregory, J.W. 1913. *The Nature and Origin of Fiords*. John Murray.

Gudlaugsson, S.T. 1993. Large impact crater in the Barents Sea. *Geology* 21(4): 291–294.

Gupta, S. et al. 2007. Catastrophic flooding origin of shelf valley systems in the English Channel. *Nature* 448: 342–345.

Gupta, S. et al. 2017. Two-stage opening of the Dover Strait. *Nature Communications* 8: 15101.

Hallet, B., Hunter, L. & Bogen, J. 1996. Rates of erosion and sediment evacuation by glaciers. *Global & Planetary Change* 12: 213–235.

Häussermann, V. et al. 2021. Species that fly at a higher game: Patterns of deep-water emergence, including a global review. *Frontiers in Marine Science* 8: 688316.

Hawkes, A.D. et al. 2007. Sediments deposited by the 2004 Indian Ocean tsunami along the Malaysia–Thailand Peninsula. *Marine Geology* 242: 169–190.

Harkrider, D. & Press, F. 1967. The Krakatoa air-sea waves: an example of pulse propagation in coupled systems. *Geophysical J. Royal Astronomical Society* 13: 149–159.

Heezen, B.C. & Ewing, M. 1963. The Mid-Oceanic Ridge. In *The Sea*, vol. 3, ed. M.N. Hill, 388–410. Interscience.

Herman, F. et al. 2015. Erosion by an Alpine glacier. *Science* 350(6257): 193–195.

Henrichs, S.M. 1992. Early diagenesis of organic matter in marine sediments: progress and perplexity. *Marine Chemistry* 39: 119–149.

Hjelstuen, B.O., Haflidason, H., Sejrup, H.P. & Lyså, A. 2009. Sedimentary processes and depositional environments in glaciated fjord systems — Evidence from Nordfjord, Norway. Marine Geology 258: 88–99.

Holliday, V.T. et al. 2023. Comprehensive refutation of the YDIH. *Earth-Science Reviews* 247: 104502.

Holtedahl, H. 1959. Geology and paleontology of Norwegian Sea bottom cores. *Journal of Sedimentary Research* 29(1): 16–29.

Holtedahl, H. 1967. Notes on the formation of fjords. *Geografiska Annaler* 49A: 188–203.

Hurst, A. A., Anderson, R. S., & Crimaldi, J. P. (2021). Toward entrainment thresholds in fluvial plucking. *Journal of Geophysical Research: Earth Surface*, 126, e2020JF005944.

Kaye, C.A. 1967. Kaolinization of bedrock of the Boston, Massachusetts area. *USGS Professional Paper* 575-C: C165–C172.

Kaye, C.A. 1984. Boston Basin restudied. In *Geology of the Coastal Lowlands, Boston to Kennebunk, Maine*, ed. L.S. Hanson, 124–140.

Kennett, J.P. et al. 2015. Bayesian chronological analyses for the YDB. *PNAS* 112(32): E4344–E4353.

Kharif, C. & Pelinovsky, E. 2005. Asteroid impact tsunamis. *Comptes Rendus Physique* 6: 361–366.

Kristoffersen, Jon Bent & Salvanes, Anne. (2009). Distribution, growth, and population genetics of the glacier lanternfish (*Benthosema glaciale*) in Norwegian waters: Contrasting patterns in fjords and the ocean. *Marine Biology Research.* 5. 596-604.

La Forge, L. 1932. Geology of the Boston area, Massachusetts. USGS Bulletin 839.

Lamb, M.P. & Fonstad, M.A. 2010. Rapid formation of a bedrock canyon by a single flood. *Nature Geoscience* 3: 477–481.

Lamb, M.P. et al. 2014. Amphitheater-headed canyons formed by megaflooding. *PNAS* 111(1): 57–62.

Larsen, E., Eide, F., Longva, O. & Mangerud, J. 1984. Allerød–Younger Dryas climatic inferences from cirque glaciers in the Nordfjord area, western Norway. *Arctic and Alpine Research* 16(2): 137–160.

Le Goff-Vitry, M.C. & Rogers, A.D. 2005. Molecular ecology of *Lophelia pertusa* in the NE Atlantic. In *Cold-Water Corals and Ecosystems*, eds. Freiwald & Roberts, 653–662. Springer.

Le Roux, J.P., et al. 2008. A Pliocene mega-tsunami deposit and associated features in the Ranquil Formation, southern Chile. *Sedimentary Geology* 203: 164–180.

Lo, C.H., et al. 2002. Laser fusion argon-40/argon-39 ages of Darwin impact glass. *Meteoritics & Planetary Science* 37: 1555–1562.

López Correa, M. et al. 2012. Preboreal onset of cold-water coral growth beyond the Arctic Circle revealed by coupled radiocarbon and U-series dating. *Quaternary Science Reviews* 34: 24–43.

MacAyeal, D.R. et al. 2006. Transoceanic wave propagation links iceberg calving margins of Antarctica with storms in tropics and Northern Hemisphere. *Geophysical Research Letters* 33: L17502.

Mangerud, J., Larsen, E., Longva, O. & Sønstegaard, E. 1979. Glacial history of western Norway 15,000–10,000 B.P. *Boreas* 8: 179–187.

Mangerud, J. et al. 2016. A major re-growth of the Scandinavian Ice Sheet in western Norway during Allerød–Younger Dryas. *Quaternary Science Reviews* 132: 175–205.

Mamo, B., Strotz, L. & Dominey-Howes, D. 2009. Tsunami sediments and their foraminiferal assemblages. *Earth-Science Reviews* 96: 263–278.

MAREANO. 2014. Geological seabed mapping along the Finnmark coast, northern Norway. Norwegian Institute of Marine Research / Geological Survey of Norway / Norwegian Hydrographic Service.

Molodtsova, T.N., Sanamyan, N.P. & Keller, N.B. 2008. Anthozoa from the northern Mid-Atlantic Ridge and Charlie-Gibbs Fracture Zone. *Marine Biology Research* 4: 112–130.

Moore, C.R. et al. 2024. Platinum, shock-fractured quartz in Eastern USA. *Airbursts & Cratering Impacts* 2(1). DOI: 10.14293/ACI.2024.0003.

Moore, C.R. et al. 2025. A 12,800-year-old layer with cometary dust, microspherules, and platinum anomaly recorded in multiple cores from Baffin Bay. *PLOS One* 20(8): e0328347.

Nalin, R., Kotulla, M., 2018. Rapid Bedrock Incision by Water Stream Outburst: The Case of the Oroville Dam (California, USA). Geoscience Research Institute. URL https://www.grisda.org/rapid-bedrock-incision-by-water-stream-outburst-the-case-of-the-oroville-dam-california-usa-1

NASA Earth Observatory. 2025. The Drumlin Islands of Boston Harbor. [Image of the Day, based on Landsat 8 imagery, July 19, 2024.]

Nathan, S. 2009. Meteorites – Meteorites in New Zealand. Te Ara – the Encyclopedia of New Zealand. https://teara.govt.nz/en/meteorites

National Park Service. 2017. NPS Geodiversity Atlas —

Boston Harbor Islands National Recreation Area, Massachusetts.

Nesje, A. et al. 1992. Quaternary erosion in the Sognefjord drainage basin. *Geomorphology* 5: 511–520.

Nesje, A. & Whillans, I.M. 1994. Erosion of Sognefjord, Norway. *Geomorphology* 9: 33–45.

Nicholson, R. 1963. A note on the relation of rock fracture and fjord direction. *Geografiska Annaler* 45: 303–304.

Nicholson, U., Jonge-Anderson, I.d., Gillespie, A. *et al.* Multiple lines of evidence for a hypervelocity impact origin for the Silverpit Crater. *Nat Commun* **16**, 8312 (2025).

Nicholson, U., et al. 2022. The Nadir Crater offshore West Africa: a candidate Cretaceous-Paleogene impact structure. *Science Advances* 8: eabn3096.

Nicholson, U., et al. 2024. 3D anatomy of the Cretaceous–Paleogene age Nadir Crater. *Communications Earth & Environment* 5: 547.

Noret, C., Girard, J. C., Munodawafa, M. C., & Mazvidza, D. Z. (2013). Kariba dam on Zambezi river: stabilizing the natural plunge pool. *La Houille Blanche*, *99*(1), 34-41.

Noret, C, Girird, J, Munodawafa, M & Mazvidza, D 2012. Kariba Dam on Zambezi River: Stabilizing the natural plunge pool. Proceedings, ICSE 6-265, Paris

O'Connor, J.E. 1993. Hydrology of the Bonneville flood. *GSA Special Paper* 274.

Ottesen, D., Dowdeswell, J.A. & Rise, L. 2005. Submarine landforms and the reconstruction of fast-flowing ice streams within a large Quaternary ice sheet: The 2500-km-long Norwegian-Svalbard margin (57°–80°N). Geological Society of America Bulletin 117: 1033–1050.

Pan, Yii-Wen & Li, Kuo-Wei & Liao, Jyh-Jong. (2014). Mechanics and Response of a Surface Rock Block Subjected to Pressure Fluctuations: A Plucking Model and Its Application. *Engineering Geology*. 171.

Pararas-Carayannis, G. 2003. Near and far-field effects of tsunamis generated by the paroxysmal eruptions of the Krakatau Volcano. *Science of Tsunami Hazards* 21(4): 191–201.

Patton, W.W. & Miller, T.P. 1978. Meteorite impact crater in central Alaska. *Science* 201(4352): 279.

Pells, Steven & Pells, Philip & Peirson, William & Douglas, K. & Fell, Robin. (2015). Erosion Of Unlined Spillways In Rock - does a "scour threshold" exist? *Contemporary Challenges for Dams;* ANCOLD, Brisbane.

Pells, Philip & Pells, Steven & Schalkwyk, A. (2016). A tale of two spillways. Appropriate technology to ensure proper Development, Operation and Maintenance of Dams in Developing Countries, *International Congress on Large Dams,* Johannesburg.

Powell, J.L. 2022. Premature rejection in science: The YDIH. *Scientific Progress* 105(1).

Pilarczyk, J.E. et al. 2020. Constraining sediment provenance for tsunami deposits using foraminiferal assemblages. *Marine Geology* 423: 106139.

Pino, M., et al. 2019. Sedimentary record from Patagonia, southern Chile supports cosmic-impact triggering of biomass burning, climate change, and megafaunal extinctions at 12.8 ka. Scientific Reports 9: 4413.

Randall, B.A.O. 1961. On the relationship of valley and fjord direction to the fracture pattern of Lyngen, Troms, N. Norway. Geografiska Annaler 43: 336–338.

Range, M.M. et al. 2022. The Chicxulub impact produced a powerful global tsunami. *AGU Advances* 3: e2021AV000627.

Rahm, D.A. 1962. Geology of the Main Drainage Tunnel, Boston, Massachusetts. *J. Boston Society of Civil Engineers* 49(3): 319–368.

Riis, F. et al. 2005. Formation of large, crater-like evacuation structures in ooze sediments in the Norwegian Sea. *Marine & Petroleum Geology* 22: 257–273.

Rosen, P. & Fitzgerald, D. (2004), Processes and Evolution of Boston Harbor Islands: Peddocks and Lovells Islands, New England Intercollegiate Geological Conference, 96th Annual Meeting, Boston, MA.

Ross, M.E. 1990. Mafic dikes of the Boston Basin area. In *Geology of the Composite Avalon Terrane of Southern New England*, eds. Socci, Skehan & Smith, GSA Special Paper 245: 133–146.

Sieh, K., et al. 2019. Australasian impact crater buried under the Bolaven volcanic field, Southern Laos. *Proceedings of the National Academy of Sciences* 117(3): 1346–1353.

Sigmond, E.M.O., Gustavson, M. & Roberts, D. 1984. Bedrock Geology Map over Norway. *Geological Survey of Norway.*

Sollid, J.L. & Sørbel, L. 1979. Deglaciation of western central Norway. *Boreas* 8: 233–239.

Smith, R.W. et al. 2015. High rates of organic carbon burial in fjord sediments globally. *Nature Geoscience* 8: 450–453.

Schmidt, R.M. & Holsapple, K.A. 1982. Estimates of crater size for large-body impact: Gravity-scaling results. In Geological Implications of Impacts of Large Asteroids and Comets on the Earth, eds. Silver & Schultz, *GSA Special* Paper 190: 93–102.

Schmidt, S., Hawkins, D., & Phillips, K. (2017, February 13). 188,000 evacuated as California's massive Oroville Dam threatens catastrophic floods. Washington Post. https://www.washingtonpost.com/news/morning-mix/wp/2017/02/13/not-a-drill-thousandsevacuated-in-calif-as-oroville-dam-threatens-to-flood/

Steer, P., Huismans, R.S., Valla, P.G., Gac, S. & Herman, F. 2012. Bimodal Plio-Quaternary glacial erosion of fjords and low-relief surfaces in Scandinavia. Nature Geoscience 5: 635–639.

Stramska, M. et al. 2018. Observations of coastal ocean currents in the Barents Sea (Porsangerfjord). *Estuarine, Coastal and Shelf Science* 200: 22–38.

Stratford, C., Lesleighter, E., Bollaert, E.. (2013). Plunge Pool Rock Scour Experiences and Analysis Techniques. IAHR Congress, Chengdu.

Stucky de Quay, G., et al. 2021. Late Holocene canyon-carving floods in northern Iceland were smaller than previously reported. *Communications Earth & Environment* 2: 164.

Syvitski, J.P.M., Burrell, D.C. & Skei, J.M. 1987. Fjords: Processes and Products. Springer-Verlag.

Sweatman, M.B. 2021. The YDIH: Review of the impact evidence. *Earth-Science Reviews* 218: 103677.

Sweatman, Martin B., James L. Powell and Allen West. "Rejection of Holliday et al.'s alleged refutation of the Younger Dryas impact hypothesis." *Earth-Science Reviews* (2024).

Suul, J., Sønstebø, G. & Nordgulen, Ø. (eds.) 2004. The West Norwegian Fjords. Norwegian Nomination 2004. UNESCO World Heritage List. Directorate for Nature Management, Ministry of the Environment, Oslo.

Tuit, C. e. (2000). Anthropogenic platinum and palladium in the sediments of Boston Harbor. *Environmental Science & Technology*.

Thompson, M.D. 2020. New insights into the Cambridge Argillite and associated intrusions, Boston Basin. In *From the Blue Hills to the Berkshires: Current Research on the Geology of Massachusetts*, ed. H.N. Berry IV, GSA Field Guide 58: 41–60. DOI: 10.1130/2020.0058(03).

Tsikalas, F., Gudlaugsson, S.T. & Faleide, J.I. 1999. The anatomy of a buried complex impact structure: The Mjølnir Structure, Barents Sea. *Journal of Geophysical Research* 103(B12): 30469–30483.

Vasskog, K. et al. 2013. Evidence for Storegga tsunami run-up at Nordfjord. *J. Quaternary Science* 28(4): 391–402.

Vorren, T.O., Hald, M. & Thomsen, E. 1984. Quaternary sediments and environments on the continental shelf off northern Norway. *Marine Geology* 57: 229–257.

Ward, S.N. 2003. Asteroid impact tsunami of 2880 March 16. *Geophysical J. International* 153: F6–F10.

Ward, S.N. & Asphaug, E. 2002. Impact tsunami—Eltanin. *Deep Sea Research* II 49: 1073–1079.

Whipple, K.X. et al. 2000. River incision into bedrock: Plucking, abrasion, and cavitation. *GSA Bulletin* 112(3): 490–503.

Wolbach, W.S. et al. 2018a. Extraordinary biomass-burning, Part 1: Ice cores. *J. Geology* 126(2): 165–184.

Wolbach, W.S. et al. 2018b. Extraordinary biomass-burning, Part 2: Sediments. *J. Geology* 126(2): 185–205.

Wünnemann, K. et al. 2006. Numerical modelling of impact tsunamis. *Geophysical J. International* 167(1): 77–88.

Wünnemann, K. & Weiss, R. 2015. The meteorite impact-induced tsunami hazard. *Phil. Trans. Royal Society A* 373: 20140381.