doi:10.1038/nature11528

Recent changes to the Gulf Stream causing widespread gas hydrate destabilization

Benjamin J. Phrampus[1] & Matthew J. Hornbach[1]

The Gulf Stream is an ocean current that modulates climate in the Northern Hemisphere by transporting warm waters from the Gulf of Mexico into the North Atlantic and Arctic oceans[1,2]. A changing Gulf Stream has the potential to thaw and convert hundreds of gigatonnes of frozen methane hydrate trapped below the sea floor into methane gas, increasing the risk of slope failure and methane release[3–9]. How the Gulf Stream changes with time and what effect these changes have on methane hydrate stability is unclear. Here, using seismic data combined with thermal models, we show that recent changes in intermediate-depth ocean temperature associated with the Gulf Stream are rapidly destabilizing methane hydrate along a broad swathe of the North American margin. The area of active hydrate destabilization covers at least 10,000 square kilometres of the United States eastern margin, and occurs in a region prone to kilometre-scale slope failures. Previous hypothetical studies[3,5] postulated that an increase of five degrees Celsius in intermediate-depth ocean temperatures could release enough methane to explain extreme global warming events like the Palaeocene–Eocene thermal maximum (PETM) and trigger widespread ocean acidification[7]. Our analysis suggests that changes in Gulf Stream flow or temperature within the past 5,000 years or so are warming the western North Atlantic margin by up to eight degrees Celsius and are now triggering the destabilization of 2.5 gigatonnes of methane hydrate (about 0.2 per cent of that required to cause the PETM). This destabilization extends along hundreds of kilometres of the margin and may continue for centuries. It is unlikely that the western North Atlantic margin is the only area experiencing changing ocean currents[10–12]; our estimate of 2.5 gigatonnes of destabilizing methane hydrate may therefore represent only a fraction of the methane hydrate currently destabilizing globally. The transport from ocean to atmosphere of any methane released—and thus its impact on climate—remains uncertain.

Methane hydrate, a solid consisting of methane and water, is stable at high pressures and low temperatures. Owing to a positive thermal gradient in the Earth, methane hydrate exists only within the first few hundred metres of sediments in deep marine settings, below which methane gas and liquid water are stable[13]. Methane hydrates represent one of the largest reservoirs of organic carbon on Earth[6,13,14]. Studies speculate that destabilization of methane hydrates could inject significant quantities of methane into the ocean and possibly the atmosphere, leading to spikes in atmospheric carbon levels[4,6,8].

The base of hydrate stability can sometimes be detected directly in seismic data via bottom-simulating reflectors (BSRs) that appear as strong, negative-polarity, high-impedance seismic reflections caused by free gas at the base of the phase boundary[15,16] (Fig. 1b). Hydrate formation is strongly dependent on temperature, and because of this, the base of the hydrate stability zone, as indicated by BSRs, is used for estimating subsurface temperature[17].

Warming waters in the Gulf Stream can potentially destabilize methane hydrate[3]. Additionally, slight changes in the Gulf Stream flow direction can also destabilize methane hydrate by introducing warm waters to regions previously exposed only to cold bottom-water currents. The Gulf Stream consists of anomalously warm water at depths as great as 1,000 metres below sea level (m.b.s.l.; Fig. 2a). In regions where the Gulf Stream is absent, ocean temperatures are markedly colder at intermediate water depths (300–1,000 m.b.s.l.; Fig. 2a).

Although much is known about the current state of the Gulf Stream, relatively little is known about the evolution of the Gulf Stream before

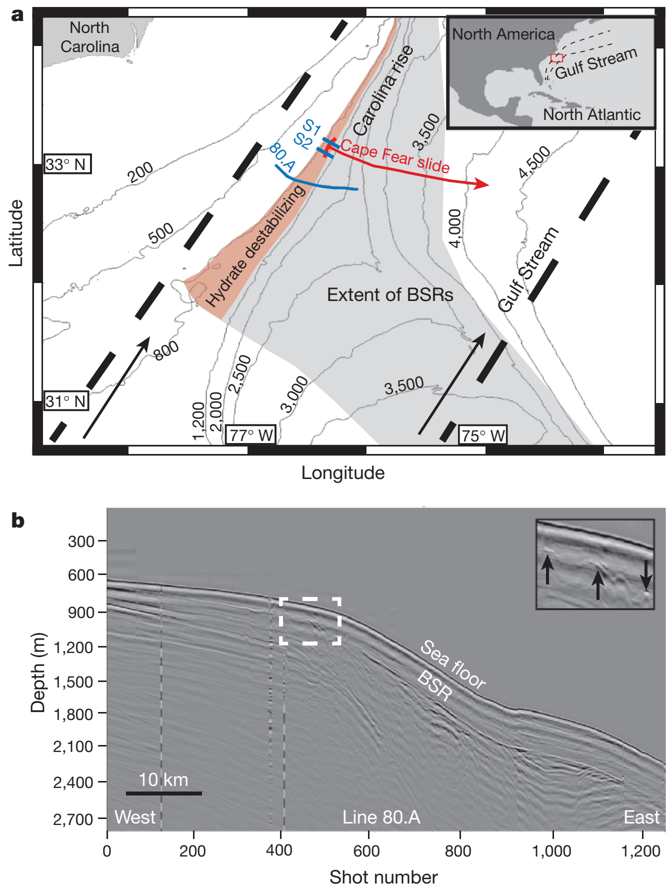

Figure 1 | Gas hydrates below the Gulf Stream in the western North Atlantic. **a**, Study area, shown boxed in the map-view inset, where the location of the Gulf Stream is also shown (dashed black lines), flowing along the western edge of the North Atlantic margin[1,2]. In the main figure, the grey area denotes where BSRs exist below the sea floor[21]; the pink area is where methane hydrate is destabilizing owing to recent changes in ocean temperature; and the approximate location of the Gulf stream is between the two solid black arrows. **b**, Multi-channel seismic line 80.A is one of several seismic lines in the region showing clear BSRs shoaling westward along the edge of the continental margin[29]. The rectangle indicated in white is shown magnified in the inset. Inset, the gas hydrate phase boundary (that is, BSR) shows as a strong, negative polarity reflector (black arrows) that behaves erratically with depth beneath sea-floor less than ~1,000 m.b.s.l.

[1]Huffington Department of Earth Sciences, Southern Methodist University, Dallas, Texas 75275, USA.

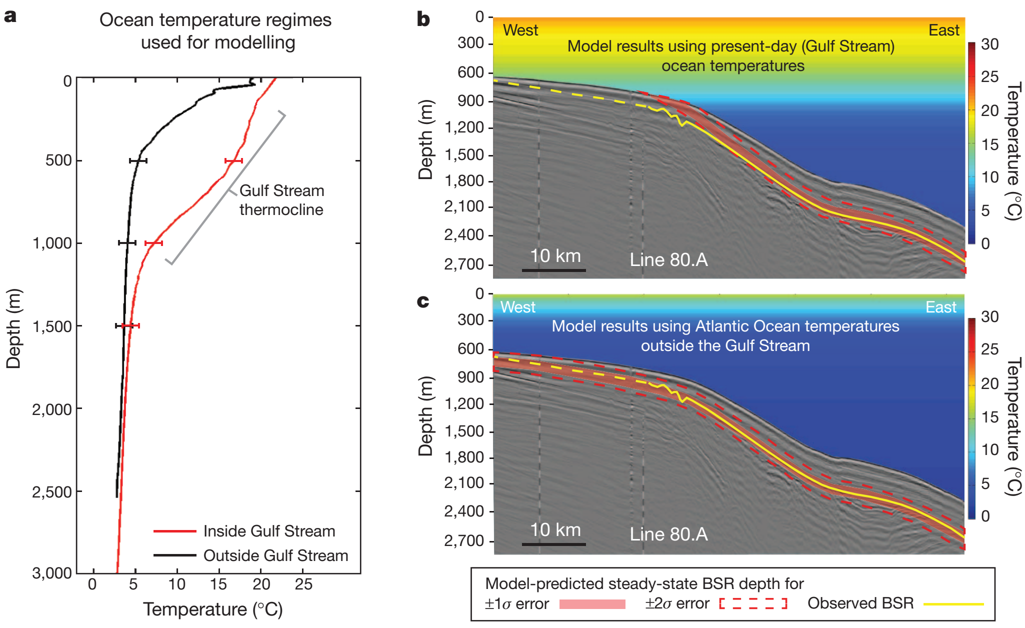

Figure 2 | Evidence for recent changes in ocean temperature and hydrate destabilization. **a**, Two different ocean temperature regimes exist in the western North Atlantic, with Gulf Stream temperatures as much as $8 \pm 1.1\,^{\circ}\mathrm{C}$ (1σ error) warmer at intermediate water depths than ocean temperatures outside the Gulf Stream. Seismic line 80.A was acquired within the Gulf Stream, yet the predicted BSR depth using present day (Gulf Stream) ocean temperatures is shallower than observed BSR depths for sea-floor depths less than ~1,000 m.b.s.l. (**b**). Comparison between observed and predicted BSR depths for the two ocean temperature regimes indicate cold, non-Gulf Stream intermediate ocean temperatures produce a much better fit to observed BSR depths (**c**). This implies recent ocean warming or northwest intrusion of the Gulf Stream along the Carolina rise and the onset of methane hydrate destabilization along the margin. This observation is widespread (Supplementary Information). Dashed yellow lines indicate where the BSR is inferred but difficult to identify clearly in seismic data.

the Holocene epoch. Lynch-Stieglitz *et al.*[1,2] and Lund *et al.*[18] suggest a lower-flux, cooler Gulf Stream during cooler climates. Alternatively, Winguth *et al.*[3], using computer simulations, suggest that during warmer climates, warm ocean temperatures like those of the Gulf Stream prevail, resulting in methane hydrate destabilization.

Where ocean temperature is well-constrained, comparison between the predicted and the actual base of methane hydrate stability offers a valuable means of identifying recent changes in ocean temperature and sub-sea-floor temperature. Here we analyse the stability of methane hydrates along the Carolina rise off the east coast of North America using active-source seismic reflection data, with the goal of characterizing hydrate stability below the Gulf Stream. The Carolina rise is a region where both the Gulf Stream and methane hydrate stability are well-constrained[1,16,19–21] (Fig. 1). Seismic data show clearly observable BSRs imaged continuously below the sea floor at water depths of 800–4,000 m.b.s.l. throughout the region[21,22] (Fig. 1b, Supplementary Information). At water depths shallower than ~1,000 m.b.s.l., the BSR shoals owing to lower hydrostatic pressures and warmer ocean temperatures, resulting in a vertically thinner hydrate stability zone within the sediments (Fig. 1b).

By comparing the predicted base of the hydrate stability zone using current hydrological and geological conditions with the observed base of the hydrate stability zone, we determine if differences exist between the observed and predicted temperature regime. Results show that at depths greater than ~1,000 m.b.s.l., the model-predicted BSR matches observed BSR depths in seismic data (Fig. 2b). At shallower depths, however, model predictions diverge from observations, with the observed BSR consistently deeper than model predictions. Even after accounting for end-member uncertainty (to 2σ), the observed BSR is deeper than model predictions (Fig. 2b).

Many phenomena might explain the difference between observed and predicted BSR depths, including changing sea level, gas composition, fluid flow, variable sedimentation rates, changes in heat flow, seismic velocity model errors, and variations in ocean temperature. Changing sea level affects hydrate stability by altering the hydrostatic pressure regime; however, sea level in the early Holocene was lower, resulting in lower pressure and shallower BSRs. Holocene changes in sea level cannot therefore explain the anomalously deep BSRs observed today. A consistent increase in the contribution of thermogenic gas along the margin could also cause the phase boundary to deepen; however, geochemical measurements indicate natural gas concentrations are consistently 99% methane across the region[19,23,24]. Fluid flow, by transporting heat more efficiently via advection, can have a significant effect on temperature; however, the anomalously deep hydrate stability zone observed along the edge of the margin requires downward flow, which is inconsistent with regional fluid models[25]. Rapid sedimentation can also result in anomalously cool shallow sediment temperatures[26]. However, sedimentation in this region has not exceeded $30\,\mathrm{m\,Myr^{-1}}$ since the Pleistocene epoch, and modelling indicates that the regional sedimentation rate is too low to explain the discrepancy between observed and predicted BSR depth[23,26]. Lateral variations in heat production can also vary heat flow and temperature. Previous studies across this region suggest no significant variations in heat flow or radiogenic material across the site[19,20,22]. Indeed, an average heat flow of $19\,\mathrm{mW\,m^{-2}}$ is necessary to explain the anomalously deep BSR at the margin edge. This is a factor of two below regional observations[20]. Finally, anomalously high seismic velocities used to create the seismic depth section could also result in anomalously deep BSRs. The seismic velocities necessary to offset observed BSR depths to model predictions are physically unreasonable[16]; a compression-wave velocity averaging less than $250\,\mathrm{m\,s^{-1}}$ in sediments above the BSR is needed to explain the discrepancy. This leaves ocean temperature variations as the last plausible explanation for the discrepancy between the observed and predicted BSR. Specifically, if ocean temperatures recently warmed at intermediate water depths, it would take time for heat to propagate downward. Before thermal equilibration, the observed BSR would therefore be cooler and appear anomalously deep compared to model predictions.

To test whether cooler intermediate-depth ocean temperatures in the recent past might explain the discrepancy between the observed and predicted depth of methane hydrate stability along the Carolina rise, we re-ran the heat-flow model but removed the warming effects of the Gulf Stream. Instead, we apply as our top boundary condition a mid-Atlantic temperature–depth profile acquired outside the Gulf Stream (Fig. 2a). Model results show a remarkably good fit between observed and predicted BSR depth for all depths (Fig. 2c).

Warming of the intermediate-depth ocean within the Holocene provides a simple explanation for the discrepancy between the observed and predicted base of the methane hydrate phase boundary (Fig. 3). The analysis implies that the Gulf Stream was cooler, lower flux, or recently diverted northwest during the Holocene, consistent with previous studies[1,2,10,18]. The result also implies that recent changes in ocean temperature are causing contemporary destabilization of methane hydrates along the North American margin.

Analysis of BSRs in other regional seismic data suggests that recent ocean bottom warming is ubiquitous along the Carolina rise. In addition to line 80.A, we analysed two other published seismic lines acquired in the Gulf Stream along the North Atlantic margin that show clearly identifiable BSRs[22] (Supplementary Information). Both lines are located more than 40–50 km away from line 80.A, yet results from all three seismic lines are identical. Specifically, analysis of all three seismic lines suggests that BSRs located below sea-floor depths shallower than ~1,000 m.b.s.l. are unstable along the Carolina rise, and that recent intermediate ocean temperature warming is widespread along the margin just east of North Carolina.

Other studies support the idea that shifting ocean currents and significant (>1 °C) Holocene ocean temperature changes occur in this region[10–12]. During the twentieth century, the most significant measured sea surface temperature warming in the North Atlantic Ocean occurred above the Carolina rise[10]. Ocean surface warming above the Carolina rise has been attributed[10] to either a northwest shift in the Gulf Stream flow direction or a warming Gulf Stream, as we suggest. Analysis of ocean drilling data also indicates recent (Holocene) changes in the Deep Western Boundary Current along the edge of this margin[11].

Extrapolating our observations across the area where the Gulf Stream and known methane hydrate deposits exist, we place first-order constraints on the area of hydrate destabilization (Fig. 1). Our analysis suggests that we are observing the onset of methane hydrate destabilization along a ~300-km span of the North American margin that will continue for centuries unless the Gulf Stream shifts southward or intermediate ocean temperatures cool several degrees[1,2,16,21] (Fig. 3). We estimate that hydrates are currently destabilizing within a volume of ~7.5×10^{11} m^3. Assuming an average porosity of 60% in the shallow sediments where hydrates are destabilizing, hydrate filling 5% of the pore space, and 123 kg of methane per cubic metre of hydrate, we estimate that ~2.5 Gt of methane—or ~0.2% of that necessary to explain the PETM—are currently destabilizing beneath a sea-floor area of ~10,000 km^2 off the US eastern seaboard[4]. If continuing hydrate destabilization triggers slope failure at this site, the amount of methane released could be an order of magnitude greater[16]. Hydrate destabilization, by converting solid methane hydrate into methane gas and water, can elevate pore fluid pressure along the slope edge, reducing slope stability in areas prone to frequent, potentially tsunamigenic slope failure[4,5,7,16]. It is perhaps no coincidence that the upper headwall of the Cape Fear slide, the largest submarine slide complex along the western North Atlantic margin, resides within the area of active hydrate destabilization (Fig. 1).

Recent shifts in Gulf Stream flow or temperature provide a simple yet powerful mechanism for contemporary methane hydrate dissociation and carbon release. The analysis presented here provides a method for constraining Holocene changes in intermediate-depth ocean temperatures and also demonstrates that slight deviations in ocean currents have a profound impact on margin stability and the ocean carbon budget. It is unlikely that the western North Atlantic margin is the only area experiencing widespread hydrate destabilization due to changing ocean currents. Recent studies have suggested that similar ocean temperature shifts may occur both in the Arctic Ocean[7,27,28] and globally along subtropical western boundary currents[12]. Our estimate of 2.5 Gt of destabilizing methane hydrate may therefore represent only a fraction of the methane hydrate currently destabilizing globally.

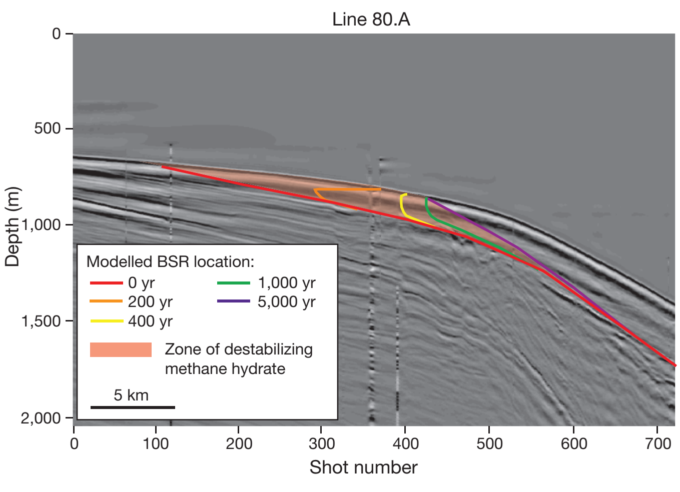

Figure 3 | Timing methane hydrate dissociation along the western Atlantic margin. To place first-order constraints on the timing of ocean temperature warming and methane hydrate dissociation along line 80.A, we run a time dependent diffusive heat-flow forward model in which we start initially with our steady-state cold ocean temperature model result (Fig. 2c), but instantaneously change the ocean temperature boundary condition to present day (Gulf Stream) values. The resulting forward model shows evolution of hydrate stability and BSR depth through time. Each coloured line represents a model-predicted BSR location in the future. These results indicate that dissociation will continue for the next several thousand years if no change in ocean temperature occurs. The pink zone represents the total area of destabilizing methane hydrate.

METHODS SUMMARY

To test whether methane hydrates below the Carolina rise are stable and temperatures are in steady state, we developed a 2D finite difference diffusive heat-flow model, discretized into 20 m (vertical) × 50 m (horizontal) cells, that accounts for conductivity, temperature, bathymetry and depth along depth-converted seismic line 80.A[29,30]. To constrain ocean temperature with depth, we average nine conductivity–temperature–depth (CTD) measurements near the site (Fig. 2a). We estimate heat flow at this site using the method outlined in ref. 17 that calculates heat flow from deep-water (>2,000 m.b.s.l.) BSRs located below the Gulf Stream thermocline. For thermal conductivity, we use a value of 1 W m^{-1} K^{-1}, consistent with regional measurements[19,20,23]. We calculate an average heat flow of 40.7 ± 4.6 mW m^{-2} (2σ), consistent with regional conductive heat-flow measurements and borehole thermal data collected near this site[19,20,23].

With conductivity, sea-floor temperature, sea-floor shape and heat flow constrained, we calculate the 2D steady-state temperature distribution within the marine sediments using a finite-difference diffusive heat flow approach. We calculate thermal diffusivity assuming constant thermal conductivity of 1 W m^{-1} K^{-1}, an average bulk density of 1,700 kg m^{-3}, and a specific heat capacity of 2,500 J K^{-1} kg^{-1}, consistent with drilling results[19,20,23]. We hold heat flow constant with time at the bottom boundary and sea-floor temperature constant with time and variable with depth, consistent with CTD values. We estimate temperature uncertainty by integrating end-member uncertainty values into the model. We calculate the base of the hydrate stability zone and location of the BSR by integrating heat flow results with the CSMGem hydrate stability program[24] assuming hydrostatic conditions and a constant salinity of 34.9‰, as measured across the region[23].

Full Methods and any associated references are available in the online version of the paper.

Received 25 May; accepted 20 August 2012.

1. Lynch-Stieglitz, J., Curry, W. & Slowey, N. Weaker Gulf Stream in the Florida Straits during the Last Glacial Maximum. *Nature* **402,** 644–648 (1999).

2. Lynch-Stieglitz, J., Schmidt, M. W. & Curry, W. Evidence from the Florida Straits for Younger Dryas ocean circulation changes. *Paleoceanography* **26,** PA1205, http://dx.doi.org/abs/10.1029/2010PA002032 (2011).
3. Winguth, A., Shellito, C., Shields, C. & Winguth, C. Climate response at the Paleocene-Eocene Thermal Maximum to greenhouse gas forcing — a model study with CCSM3. *J. Clim.* **23,** 2562–2584 (2010).
4. Dickens, G. R. The potential volume of oceanic methane hydrates with variable external conditions. *Org. Geochem.* **32,** 1179–1193 (2001).
5. Hornbach, M. J., Saffer, D. M. & Holbrook, W. S. Critically pressured free-gas reservoirs below gas-hydrate provinces. *Nature* **427,** 142–144 (2004).
6. Kennett, J. P., Cannariato, K. G., Hendy, I. L. & Behl, R. J. *Methane Hydrates in Quaternary Climate Change: The Clathrate Gun Hypothesis* (Spec. Publ. Ser. Vol. 54, American Geophysical Union, 2003).
7. Biastoch, A. *et al.* Rising Arctic Ocean temperatures cause gas hydrate destabilization and ocean acidification. *Geophys. Res. Lett.* **38,** L08602, http://dx.doi.org/abs/10.1029/2011GL047222 (2011).
8. Archer, D. *et al.* The importance of ocean temperature to global biogeochemistry. *Earth Planet. Sci. Lett.* **222,** 333–348 (2004).
9. Flemings, B. P., Liu, X. & Winters, W. J. Critical pressure and multiphase flow in Blake Ridge gas hydrates. *Geology* **31,** 1057–1060 (2003).
10. Deser, C. & Blackmon, M. L. Surface climate variations over the North Atlantic Ocean during winter: 1900–1989. *J. Clim.* **6,** 1743–1753 (1993).
11. Evans, H. K., Hall, I. R., Bianchi, G. G. & Oppo, D. W. Intermediate water links to Deep Western Boundary Current variability in the subtropical NW Atlantic during marine isotope stages 5 and 4. *Paleoceanography* **22,** PA3209, http://dx.doi.org/abs/10.1029/2006PA001409 (2007).
12. Wu, L. *et al.* Enhanced warming over the global subtropical western boundary currents. *Nature Clim. Change* **2,** 161–166 (2012).
13. Kvenvolden, K. A. Gas hydrates: geological perspective and global change. *Rev. Geophys.* **31,** 173–187 (1993).
14. Milton, D. J. in *The Future Supply of Nature-Made Gas* (ed. Meyer, R. F.) 927–943 (Technical Reports, Pergamon, 1976).
15. Shipley, T. H. *et al.* Seismic evidence for widespread possible gas hydrate horizons on continental slopes and rises. *Bull. Am. Assoc. Petrol. Geol.* **63,** 2204–2213 (1979).
16. Holbrook, W. S., Hoskins, H., Wood, W. T., Stephen, R. A. & Lizarralde, D. Methane hydrate and free gas on the Blake Ridge from vertical seismic profiling. *Science* **273,** 1840–1843 (1996).
17. Yamano, M., Uyeda, S., Aoki, Y. & Shipley, T. H. Estimates of heat flow derived from gas hydrates. *Geology* **10,** 339–343 (1982).
18. Lund, D., Lynch-Stieglitz, J. & Curry, W. Gulf Stream density structure and transport during the past millennium. *Nature* **444,** 601–604 (2006).
19. Ruppel, C. D. Anomalously cold temperatures observed at the base of the gas hydrate stability zone on the US Atlantic passive margin. *Geology* **25,** 699–702 (1997).
20. Ruppel, C. D., Von Herzen, R. P. & Bonneville, A. Heat flux through an old (~175 Ma) passive margin: offshore southeastern United States. *J. Geophys. Res.* **100** (B10), 20037–20057 (1995).
21. Dillon, W. P., Lee, M. W., Fehlhaber, K. & Coleman, D. F. Gas hydrates on the Atlantic margin of the United States — controls on concentration. *US Geol. Surv. Prof. Pap.* **1570,** 313–330 (1993).
22. Carpenter, G. Coincident sediment slump/clathrate complexes on the U.S. Atlantic continental slope. *Geo-Mar. Lett.* **1,** 29–32 (1981).
23. Paull, C. K., Matsumoto, R. & Wallace, P. J. Gas hydrate sampling on the Blake Ridge and Carolina Rise. *Proc. ODP Init. Rep.* **164,** 5–12, http://dx.doi.org/10.2973/odp.proc.ir.164.1996 (1996).
24. Sloan, E. D. & Koh, C. A. *Clathrate Hydrates of Natural Gases* 3rd edn (CRC Press, 2008).
25. Dugan, B. & Fleming, P. B. Overpressure and fluid flow in the New Jersey continental slope: implications for slope failures and cold seeps. *Science* **289,** 288–291 (2000).
26. Hutchison, I. The effects of sedimentation and compaction on oceanic heat flow. *Geophys. J. R. Astron. Soc.* **82,** 439–459 (1985).
27. Westbrook, G. K. *et al.* Escape of methane gas from the seabed along the West Spitsbergen continental margin. *Geophys. Res. Let.* **36,** L15608, http://dx.doi.org/abs/10.1029/2009GL039191 (2009).
28. Rajan, A. Mienert, J. & Bünz, S. Acoustic evidence for a gas migration and release system in Arctic glaciated continental margins offshore NW-Svalbard. *Mar. Petrol. Geol.* **32,** 36–49 (2012).
29. Bryan, G. M., Markl, R. G. & Sheridan, R. E. IPOD site survey in the Blake-Bahamas Basin. *Mar. Geol.* **35,** 43–63 (1980).
30. Mackenzie, K. V. Discussion of seawater sound-speed determinations. *J. Acoust. Soc. Am.* **70,** 801–806 (1981).

Supplementary Information is available in the online version of the paper.

Acknowledgements We thank L. Gahagan and the UTIG Marine Seismic Data Center for providing access to seismic line 80.A.

Author Contributions B.J.P. performed all experimental work integrating seismic lines and CTD temperature data into 2D thermal models and the hydrate stability model. M.J.H. conceived the work and developed and tested the prototype 2D models. Both authors worked together to refine the model, analyse data, interpret results, write the manuscript and create the figures.

Author Information Reprints and permissions information is available at www.nature.com/reprints. The authors declare no competing financial interests. Readers are welcome to comment on the online version of the paper. Correspondence and requests for materials should be addressed to B.J.P. (bphrampus@mail.smu.edu).

METHODS

Data preparation. We obtained line 80.A seismic data and navigation files from the UTIG marine seismic data repository. The data, originally collected in 1977[29], were provided as a stacked time section in SEGY format. Original processing of the seismic data included demultiplexing, CDP sorting and velocity analysis using semblance techniques, normal moveout correction, 24-fold stacking, time-variable gain, and band-pass filtering. Because the data supplied are post-stack, we used velocity analysis of CDP gathers from long-offset regional pre-stack seismic data acquired near Blake ridge in 2000 to estimate seismic velocities in the shallow (<500 m.b.s.l.) subsurface. Interval velocities at depths <500 m.b.s.l. consistently average just above ~1,500 m s^{-1}, and we used these interval velocities to convert line 80.A from time to depth. Additionally, to convert the seismic section to depth, we incorporated conductivity–temperature–depth (CTD) data to account for the effects of ocean temperature and conductivity on sound speed with depth[30]. The depth conversion enabled us to estimate sea-floor and BSR depth with metre-scale accuracy.

Heat flow model. With the 2D seismic depth profile for line 80.A constrained, we create the domain for the 2D conductive heat flow model by picking the depth of the sea floor along the profile using standard seismic interpretation software. We export sea-floor picks into Matlab, and digitize them with a vertical resolution of 20 m and horizontal resolution of 50 m. This is the approximate resolution of the seismic data and represents the standard grid-size resolution we use to run the 2D heat flow model.

Next, we create a 1,500 × 255 cell thermal conductivity grid (with 50 m × 20 m cell dimensions), in which we assign a conductivity value of 1 W m^{-1} K^{-1} for all depths below the sea floor, consistent with regional measurements[19,20,23]. The base of the conductivity model is such that the basal boundary locations are greater than or equal to 3 km below the sea floor. We calculate thermal diffusivity using the conductivity grid by assuming a constant thermal conductivity of 1 W m^{-1} K^{-1}, an average density of 1,700 kg m^{-3} for the sediments and a specific heat capacity of 2,500 J K^{-1} kg^{-1}, consistent with drilling results[19,20,23].

We also create a 1,500 × 255 cell temperature grid. To define the values for the initial conditions and the boundary conditions of the temperature grid, we use previously published regional heat flow measurements as well as independent heat flow calculations from deep BSRs to constrain heat flow across line 80.A[17,19,20,24]. We calculate the regional heat flow using BSRs in the deep water environment (>2,000 m.b.s.l.) where little variability in bottom water temperature exists by assuming (1) a hydrate stability pressure–temperature curve for pure methane, (2) a salinity of 34.9‰ and (3) hydrostatic conditions. We use the CSMGem program to constrain hydrate stability for these conditions[24]. The heat flow value we obtain using this method of 40.7 ± 4.6 mW m^{-2} (2σ) is consistent with independent regional conductive heat-flow measurements and borehole thermal data collected near this site[19,20,23].

We define our initial starting-model temperature grid by first assuming a 1D linear temperature gradient (that is, temperature increasing linearly with depth below the sea floor), where we use two different ocean temperature regimes (inside the Gulf Stream versus outside the Gulf Stream) as the top boundary condition at the sea floor for our two different models. To constrain ocean temperatures at the sea floor beneath the Gulf Stream, we used nine CTD profiles located near our study area obtained from the World Ocean Circulation Experiment (WOCE). Using these sites, we found temperature versus depth below the Gulf Stream had an average standard deviation of ±1.1 °C. For sea-floor temperatures outside the Gulf Stream, we used the closest WOCE CTD measurements available near our study area outside the Gulf Stream. As an additional cold water test, we used CTDs from the tropical eastern Pacific where warm surface waters exist but no Gulf Stream exists. CTD temperature–depth profiles for the eastern Pacific and non-Gulf Stream Atlantic were consistent to within ~1.5 °C, resulting in minimal difference in model results. For each of the models shown in Fig. 2, sea-floor temperature is held constant in time but is variable with depth, consistent with measured temperature–depth profiles. With the top boundary condition constrained by CTD data, we assume open side boundary conditions, but hold the top boundary and basal boundary temperature constant in time.

With thermal conductivity and temperature at the boundary conditions constrained, we run the 2D finite difference heat flow model. The 2D model assumes a time-dependent, conductive heat flow regime with constant diffusivity, and no significant *in situ* generation of heat. It therefore uses the following equation to solve for temperature changes in time and space:

$$\frac{\partial T(x,y,t)}{\partial t} = \kappa\left(\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial y^2}\right) \tag{1}$$

Here, ∂ is a partial derivative, T is temperature, t is time, κ is the diffusivity, x is position in the x-direction, and y is position in the y-direction. To solve this numerically, we use a finite-difference technique where we approximate and expand the space derivatives to second order such that:

$$\frac{\partial^2 T^t_{x,y}}{\partial x^2} \approx \frac{\left(T^t_{x+1,y} - 2T^t_{x,y} + T^t_{x-1,y}\right)}{\Delta x^2} \tag{2}$$

and

$$\frac{\partial^2 T^t_{x,y}}{\partial y^2} \approx \frac{\left(T^t_{x,y+1} - 2T^t_{x,y} + T^t_{x,y-1}\right)}{\Delta y^2} \tag{3}$$

Where $T^t{}_{x,y}$ is the temperature at time t and cell position (x, y); $T^t{}_{x+1,y}$ is the temperature at time t and cell position $(x + 1, y)$; $T^t{}_{x-1,y}$ is the temperature at time t, and cell position $(x - 1, y)$; $T^t{}_{x,y+1}$ is the temperature at time t and cell position $(x, y + 1)$; and $T^t{}_{x,y-1}$ is the temperature at time t and cell position $(x, y - 1)$. Here, δx and δy are approximated as $\Delta x = 50$ m and $\Delta y = 20$ m, respectively, as defined by our grid size. We solve the time dependent heat flow equation using a semi-implicit (forward-time centre-space) technique, noting we that we can approximate the change in temperature with time at any given cell as:

$$\frac{\partial T(x,y,t)}{\partial t} \approx \frac{\left(T^{t+1}_{x,y} - T^t_{x,y}\right)}{\Delta t} \tag{4}$$

where $T^{t+1}{}_{x,y}$ is the temperature at cell (x, y) at the next time step $(t + 1)$, and Δt is the time step. To maintain numerical stability, we set Δt less than $\Delta y^2/(2\kappa)$. We iteratively determine how temperature changes in time and space below the sea floor at line 80.A by substituting equations (2), (3) and (4) into equation (1) and solving for T^{t+1}, such that:

$$T^{t+1}_{x,y} = T^t_{x,y} + \kappa\Delta t\left[\frac{\left(T^t_{x+1,y} - 2T^t_{x,y} + T^t_{x-1,y}\right)}{\Delta x^2} + \frac{\left(T^t_{x,y+1} - 2T^t_{x,y} + T^t_{x,y-1}\right)}{\Delta y^2}\right]$$

To calculate steady-state temperatures for each of the models presented in Fig. 2, we ran the heat flow model for 3 million years, using the two different ocean temperature regimes for each model. We estimate temperature uncertainty for Fig. 2 by re-running the model using end-member (2σ) uncertainty values for both ocean temperature and heat flow.

To calculate the expected temperature change with time shown in Fig. 3, we took the steady-state solution for the cold ocean temperature regime as our starting model but set the upper boundary condition to warm Gulf Stream temperatures values. This model therefore assumes an instantaneous ocean temperature change. We then ran the model for 10,000 years, outputting temperature results at 100-year time intervals.

CSMGem hydrate stability model. To calculate the location where the base of hydrate stability (and therefore, the BSR) exists below seismic line 80.A, we integrate our heat flow model results with the CSMGem program[24]. The CSMGem program generates precise and accurate hydrate stability phase diagrams that account for natural gas composition, concentration of inhibitors of hydrate formation (such as salt), and variable pressure and temperature regimes.

With steady-state temperature constrained below the sea floor via 2D numerical modelling results, we determine the locations of hydrate stability below the sea floor by estimating pressure at each cell in the model. For pressure, we assume both hydrostatic and lithostatic conditions. For hydrostatic conditions, we assume a water density of 1,040 kg m^{-3}; for lithostatic conditions, we assume an average grain density of 2,700 kg m^{-3}. We assign both lithostatic and hydrostatic pressure values to each cell in the temperature grid.

To use the CSMGem model to constrain where hydrate forms, we assume pure methane gas and a constant salinity of 34.9‰, in accordance with previous studies near this site[19,20,23]. We now have constraints on chemistry, temperature (from 2D model results), and pressure (hydrostatic and lithostatic), and can calculate, based on CSMGem output results, where hydrate is stable for any given pressure or temperature value.

With the hydrate stability phase boundary known from CSMGem results, we interrogate each cell location in the temperature and pressure grids to determine if the pressure and temperature conditions are appropriate for hydrate stability at each cell. Cells where hydrate is stable are denoted in a new hydrate stability grid with the value "1". At cells where hydrate is unstable, we fill the hydrate stability grid with a value of "0". The program then searches through the hydrate stability grid and auto picks the location of BSRs by finding where transitions occur between cell values of 1 and 0. The hydrate stability grid therefore produces the model results that are overlain on the seismic data in Figs 2 and 3.