GEOPHYSICAL RESEARCH LETTERS, VOL. 37, L12701, doi:10.1029/2010GL043473, 2010

Sensitivities of zonal mean atmospheric circulation to SST warming in an aqua-planet model

Gang Chen,[1] R. Alan Plumb,[2] and Jian Lu[3,4]

Received 30 March 2010; revised 5 May 2010; accepted 10 May 2010; published 16 June 2010.

[1] Sensitivities of tropospheric winds and stratospheric Brewer-Dobson Circulation (BDC) to SST warming are explored in an aqua-planet atmospheric general circulation model. The tropospheric zonal wind change is quite sensitive to the location and sign of the gradient of SST perturbations with respect to the climatological jet. For the experiments with low latitude warming, the Hadley cell is intensified in the deep tropics, yet the Hadley cell boundary contracts for narrow meridional extents of warming and expands for broad extents of warming, associated with changes in extratropical eddy-driven winds. Despite the complex changes of tropospheric wave forcing, the strength of the BDC is increased for all the experiments with low latitude warming. For the experiments with high latitude warming, the strength of the BDC decreases only if the warming extending to the subtropics. **Citation:** Chen, G., R. A. Plumb, and J. Lu (2010), Sensitivities of zonal mean atmospheric circulation to SST warming in an aqua-planet model, *Geophys. Res. Lett.*, *37*, L12701, doi:10.1029/2010GL043473.

1. Introduction

[2] There is increasing evidence of large-scale atmospheric circulation changes in recent decades as a result of climate change. Models of Intergovernmental Panel on Climate Change (IPCC)/Fourth Assessment Report, with a focus on the troposphere, show a consistent poleward shift of midlatitude jet streams, storm tracks, or the boundaries of Hadley cell circulations under global warming [e.g., *Yin*, 2005; *Miller et al.*, 2006; *Lu et al.*, 2007, 2008; *Chen and Held*, 2007; *Chen et al.*, 2008]. More recent climate models with improved representations of stratospheric physics and dynamics, unanimously predict an acceleration of the Brewer-Dobson Circulation (BDC) under climate change [e.g., *Butchart et al.*, 2006; *Li et al.*, 2008; *Garcia and Randel*, 2008; *McLandress and Shepherd*, 2009]. The strengthening of the BDC has, in turn, been attributed to increases of wave drag through the "downward control" mechanism [*Haynes et al.*, 1991].

[3] On the observational side, the poleward shift of jet streams is seen in the reanalysis winds [e.g., *Thompson et al.*, 2000] and satellite-derived tropospheric temperature record [*Fu et al.*, 2006]. Although the recent BDC increase is seemingly consistent with satellite-retrieved stratospheric temperature records in the Southern Hemisphere [*Lin et al.*, 2009], there is no evidence of trend in the age of stratospheric air derived from balloon-borne measurements of stratospheric trace gases [*Engel et al.*, 2009].

[4] What causes these circulation changes is not fully understood. Since global warming and El Niño are both characteristic of tropical upper tropospheric warming due to water vapor feedback [*Held and Soden*, 2000], one may be tempted to speculate that the two climate forcings would lead to similar circulation changes. Indeed, more planetary Rossby waves are seen in climate models to propagate into the middle atmosphere during El Niño years, driving a stronger BDC [e.g., *García-Herrera et al.*, 2006]. In the troposphere, however, the atmosphere displays nearly opposite changes in the Hadley cell width and the latitude of surface westerlies in response to El Niño versus global warming [*Lu et al.*, 2008; *Chen et al.*, 2008].

[5] Interestingly, a dry atmospheric dynamical core driven by *Held and Suarez* [1994] type of forcing shows different responses to idealized tropical warming. *Eichelberger and Hartmann* [2005] and *Son and Lee* [2005] separately showed a strengthening of the BDC and an equatorward jet shift in response to tropical warming, but *Butler et al.* [2010] found that tropical warming leads to a weakening of the BDC and a poleward jet shift. While the apparent discrepancy in different studies may be attributable to details in the model configuration, differences in the imposed heating in this type of model may account for the disagreement from more realistic climate models.

[6] In this study, we examine sensitivities of zonal mean atmospheric circulation to different patterns of surface warming in an aqua-planet moist model. In a similar model, *Brayshaw et al.* [2008] have shown that the locations of midlatitude storm tracks are sensitive to SST forcing. The focus here is the sensitivities of the location of tropospheric winds and the strength of the BDC in the stratosphere. Our results support those circulation changes found in climate models.

2. Model Characteristics

[7] We use the aqua-planet version of GFDL atmospheric model AM2.1 [*Delworth et al.*, 2006]. This is a low top model of the stratosphere: there are five levels in the stratosphere, with the top level at about 3 hPa. This is sufficient for our study, since the BDC associated with synoptic wave forcing is confined within the lower stratosphere [*Plumb*, 2002]. Our control experiment follows the idealized forcings (Qobs SST) described in the aqua-planet model benchmark [*Neale and Hoskins*, 2000]. The prescribed SST

[1] Department of Earth and Atmospheric Sciences, Cornell University, Ithaca, New York, USA.
[2] Program in Atmospheres, Oceans and Climate, Massachusetts Institute of Technology, Cambridge, Massachusetts, USA.
[3] Center for Ocean-Land-Atmosphere Studies, Calverton, Maryland, USA.
[4] Department of Atmospheric, Oceanic and Earth Sciences, George Mason University, Fairfax, Virginia, USA.

0094-8276/10/2010GL043473

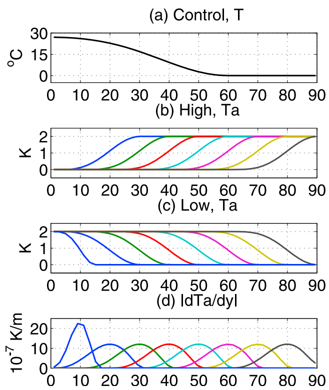

Figure 1. The SST distribution specified in the aqua-planet model. (a) The SST profile in the control experiment. (b, c) Two sets of SST perturbations with anomalous SST warming in the high latitudes and low latitudes, respectively. (d) The magnitude of meridional gradient of SST perturbations in Figure 1c. The anomalous warming decreases rapidly away from the latitudinal range of warming, and the latitude of anomalous SST gradient maximum is varied approximately from 10 to 80 degrees latitude with an increment of 10 degrees. Each pair of experiments adds up to 2K uniform warming.

profile is zonally symmetric and has a latitudinal structure as Figure 1, and there is no sea ice in the model. The solar radiative forcing is fixed in the equinoctial condition. As the prescribed SST and radiative forcings are zonally symmetric, there are no stationary waves in the model, yet there are instantaneous eddies. Since the forcings are symmetric about the equator, the time mean results are hemispherically averaged over the two hemispheres. Our results are averaged over 6 years of integration after the 1-year spin-up is discarded.

[8] The climatology in our control experiment is similar to that of *Brayshaw et al.* [2008]. There is a tropospheric jet at about 35° latitude and 200 hPa. Because of the predominant synoptic wave forcing, the simulated mean residual meridional circulation rises in the tropics and descends in the midlatitudes, as expected from the BDC branch due to synoptic eddies [*Plumb*, 2002].

[9] The SST profile is perturbed systematically in two sets of experiments with anomalous SST warming in the high latitudes and low latitudes, respectively.

$$
\begin{aligned}
Ta(\phi)_{\mathrm{High}} &= 2 \times \sin^4(3 \times \max(\min(|\phi| - \phi_0, 30^\circ), 0^\circ)) \\
Ta(\phi)_{\mathrm{Low}} &= 2 - Ta(\phi)_{\mathrm{High}}, (\phi_0 = 0^\circ, 10^\circ, \cdots, 60^\circ)
\end{aligned}
\tag{1}
$$

Figure 1 shows that the anomalous warming decreases rapidly away from the latitudinal range of warming, and the latitude of anomalous SST gradient maximum is approximately ϕ_0+20°, varying from 20 to 80 degrees latitude with an increment of 10 degrees. There is an additional experiment of low latitude warming with maximum gradient at about 10°.

$$
Ta(\phi)_{\mathrm{Low,\ Trop}} = 2 - 2 \times \sin^4(6 \times \max(\min(|\phi|, 15^\circ), 0^\circ)) \tag{2}
$$

3. Results

[10] Figure 2 shows the responses of the latitude of surface westerly maximum and the strength of stratospheric mass flux to anomalous high latitude and low latitude SST warming, respectively. Note that the results are plotted as a function of the latitude of anomalous SST gradient maximum. Despite the ideal nature of the SST forcings, key features of jet stream and BDC variability are reproduced. In response to tropical warming, the jet moves equatorward and the BDC is strengthened, analogous to the atmospheric response to El Nino [i.e., *Seager et al.*, 2003; *L'Heureux and Thompson*, 2006; *García-Herrera et al.*, 2006]. As the meridional extent of the low latitude warming is extended to the midlatitudes, the change of the BDC strength remains positive, but the direction of tropospheric jet shift changes from equatorward to poleward, analogous to the contrast between the atmospheric responses to El Nino versus global warming [*Lu et al.*, 2008; *Chen et al.*, 2008].

[11] For the high latitude warming confined poleward of 40° latitude, the BDC strength is nearly unchanged, and the midlatitude jet moves equatorward, reminiscent of the impact of high latitude forcing on the phase of NAO indi-

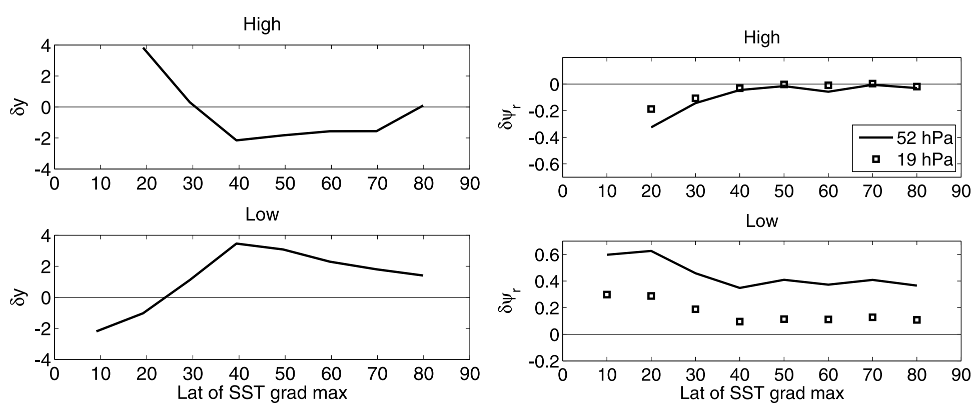

Figure 2. The changes of (left) the latitude of surface westerly maximum (indicating the location of an eddy-driven jet) and (right) the strength of stratospheric mass flux (residual mean streamfunction maximum) to (top) high latitude and (bottom) low latitude SST warming in Figure 1. The responses are displayed as a function of the latitude of anomalous SST gradient maximum. The unit for the jet latitude shift is degrees of latitude with a positive value for a poleward shift. The unit for the mass flux is 10^9 kg s^{-1} with a positive value for a strengthening of the BDC.

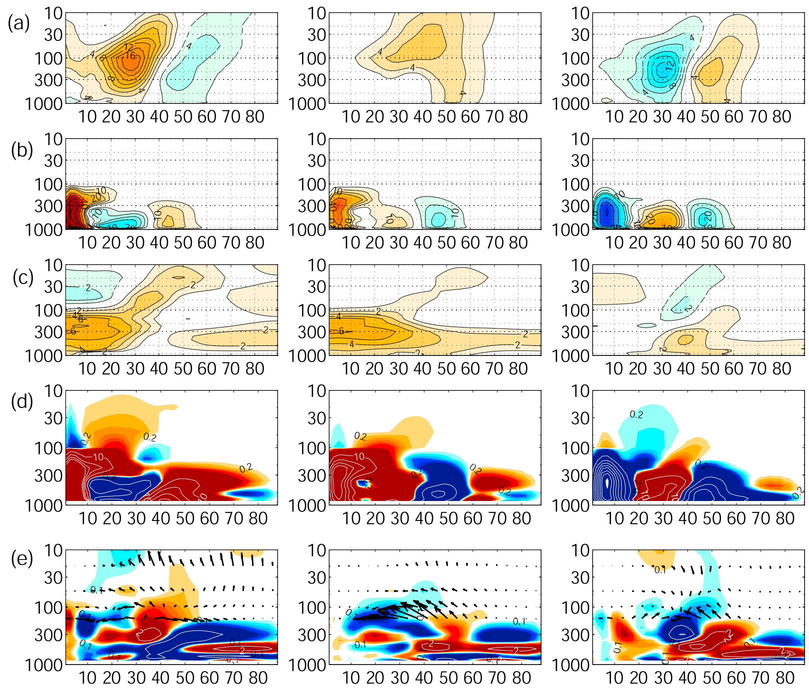

Figure 3. The circulation changes in three experiments: anomalous low latitude SST warming with maximum gradient separately at (left) 10° and (middle) 60°, and (right) anomalous high latitude warming with maximum gradient at 20°. The variables and contour intervals are: (a) zonal wind (2 m s^{-1}); (b) mean meridional streamfunction (5×10^9 kg s^{-1}); (c) temperature (1 K); (d) residual mean meridional streamfunction (0.2×10^9 kg s^{-1} for color shading and 10×10^9 kg s^{-1} for white contours); (e) EP flux divergence (0.1 m $s^{-1}$$\text{day}^{-1}$ for color shading and 2 m $s^{-1}$$\text{day}^{-1}$ for white contours) and EP vectors (arrows). Note that only the levels above 300 hPa are plotted for EP vectors.

rectly through changes in extratropical eddies [i.e., *Deser et al.*, 2004].

[12] More specifically, the tropospheric jet response is quite sensitive to the location and sign of the gradient of SST perturbations with respect to the climatological jet at about 35° latitude. For high latitude warming, the meridional temperature gradient is reduced, and the jet moves equatorward for warming confined poleward of the climatological jet. The meridional temperature gradient is increased for low latitude warming, and the jet moves poleward for the warming extending poleward of the jet latitude. Generally speaking, the jet moves towards (away from) the flank where the baroclinic wave generation is enhanced (decreased) due to changes of anomalous SST and tropospheric temperature gradient (i.e., where the Eady growth rate is enhanced (decreased)). Analogous latitudinal shifts are found for the midlatitude storm tracks in an aquaplanet model [*Brayshaw et al.*, 2008] and the annular mode-like responses to thermal forcings in a dry model [*Ring and Plumb*, 2008], since the storm tracks or jet streams are largely driven by baroclinic eddies. Despite the complexity in the change of tropospheric wave forcing, the strength of the BDC is increased for all the experiments with low latitude warming. For the experiments with high latitude warming, the strength of the BDC decreases only if the warming extending to the subtropics.

[13] We further explore these sensitivities with three experiments in Figure 3. Figures 3 (left) and 3 (middle) are the circulation responses to low latitude warming with narrow and broad meridional extents (i.e., anomalous SST gradient maximum at 10° and 60°, respectively). In both experiments, there are similar warming in tropical tropospheric temperature and intensification of the Hadley cell in the deep tropics, as expected from tropical SST warming and impacts on the lapse rate and diabatic heating. However, the meridional shift in the subtropical boundary of the Hadley cell is opposite in the two experiments, which, in turn, may be attributed to different responses of extratropical eddies accompanying changes in baroclinicity [*Ring and Plumb*, 2008], tropospheric static stability [*Lu et al.*, 2008], or tropospheric eddy phase speeds [*Chen et al.*, 2007, 2008]. This implies that the Hadley cell in the deep tropics and in the subtropics may be separately dominated by diabatic heating and extratropical eddies. Additionally in the stratosphere, the BDC is increased poleward of about 10° latitude in both cases, leading to similar midlatitude warming in the stratosphere.

[14] We examine the cause of stratospheric change by comparing the two experiments in Figures 3 (middle) and 3 (right) (i.e., low latitude warming extending to 60° and high latitude warming extending to 20°, respectively). The tropospheric jet moves poleward in both experiments, associated with similar anomalous EP flux divergence across the tropopause at about 50°. However, the changes in the strength of the BDC and associated stratospheric midlatitude temperature are opposite for the two experiments. The difference in the BDC strength can be attributed to anomalous EP flux convergence in the subtropical stratosphere by the downward control mechanism [*Haynes et al.*, 1991]. The difference in subtropical wave drag, in turn, is due to the difference in the subtropical wind. For example, tropical tropospheric warming and associated thermal wind can lead

to increased subtropical wave drag [*Garcia and Randel*, 2008]. These results suggest that the subtropical branch of the BDC associated with synoptic waves is primarily controlled by changes in the subtropical winds, which provide a waveguide to stratospheric waves to penetrate into the tropics, rather than the latitude of tropospheric waves propagating into the stratosphere.

4. Conclusions

[15] We have explored sensitivities of zonal mean circulation to different SST warming using an aqua-planet atmospheric general circulation model. Unlike previous studies with dry atmospheric models where the results depend on details of imposed tropical heating [cf. *Eichelberger and Hartmann*, 2005; *Son and Lee*, 2005; *Butler et al.*, 2010], our moist model results are consistent with realistic climate model simulations on the atmospheric responses to tropical upper tropospheric warming associated with global warming and El Nino in both the troposphere and stratosphere.

[16] Our results show that the tropospheric jet change is quite sensitive to the location and sign of the gradient of SST perturbations with respect to the climatological jet, as by *Ring and Plumb* [2008]. Roughly speaking, the jet moves towards the flank where baroclinic wave generation is enhanced due to an increase of anomalous SST gradient, and it moves away from the jet flank where baroclinicity is reduced. For the experiments of low latitude warming, the Hadley cell is intensified in the deep tropics in all the experiments, yet the Hadley cell boundary contracts for narrow warming extents and expands for broad warming extents. This is consistent with extratropical zonal wind. Despite the complexity in the change of tropospheric wave forcing, the strength of the BDC is increased for all the experiments with low latitude warming. For the experiments with high latitude warming, the strength of the BDC decreases only if the warming extending to the subtropics.

[17] Since the SST forcing imposed is zonally symmetric, our model do not have stationary planetary waves. This necessitates more works to explore the importance of zonal asymmetrical SST forcing in the atmospheric response to surface warming patterns.

[18] **Acknowledgments.** We thank two anonymous reviewers who have helped to improve the clarity of the manuscript.

References

Brayshaw, D. J., B. Hoskins, and M. Blackburn (2008), The storm-track response to idealized SST perturbations in an aquaplanet GCM, *J. Atmos. Sci.*, *65*, 2842–2860.
Butchart, N., et al. (2006), Simulations of anthropogenic change in the strength of the Brewer-Dobson circulation, *Clim. Dyn.*, *27*, 727–741, doi:10.1007/s00382-006-0162-4.
Butler, A. H., D. W. J. Thompson, and R. Heikes (2010), The steady-state atmospheric circulation response to climate change-like thermal forcings in a simple General Circulation Model, *J. Clim.*, in press.
Chen, G., and I. M. Held (2007), Phase speed spectra and the recent poleward shift of Southern Hemisphere surface westerlies, *Geophys. Res. Lett.*, *34*, L21805, doi:10.1029/2007GL031200.
Chen, G., I. M. Held, and W. A. Robinson (2007), Sensitivity of the latitude of the surface westerlies to surface friction, *J. Atmos. Sci.*, *64*, 2899–2915.
Chen, G., J. Lu, and D. M. W. Frierson (2008), Phase speed spectra and the latitude of surface westerlies: Interannual variability and global warming trend, *J. Clim.*, *21*, 5942–5959.
Delworth, T. L., et al. (2006), GFDL's CM2 global coupled climate models. Part I: Formulation and simulation characteristics, *J. Clim.*, *19*, 643–674.
Deser, C., G. Magnusdottir, R. Saravanan, and A. Phillips (2004), The effects of North Atlantic SST and sea ice anomalies on the winter circulation in CCM3. Part II: Direct and indirect components of the response, *J. Clim.*, *17*, 877–889.
Eichelberger, S. J., and D. L. Hartmann (2005), Changes in the strength of the Brewer-Dobson circulation in a simple AGCM, *Geophys. Res. Lett.*, *32*, L15807, doi:10.1029/2005GL022924.
Engel, A., et al. (2009), Age of stratospheric air unchanged within uncertainties over the past 30 years, *Nat. Geosci.*, *2*, 28–31, doi:10.1038/ngeo388.
Fu, Q., C. M. Johanson, J. M. Wallace, and T. Reichler (2006), Enhanced mid-latitude tropospheric warming in satellite measurements, *Science*, *312*, 1179, doi:10.1126/science.1125566.
Garcia, R. R., and W. J. Randel (2008), Acceleration of the Brewer-Dobson circulation due to increases in greenhouse gases, *J. Atmos. Sci.*, *65*, 2731–2739.
García-Herrera, R., N. Calvo, R. R. Garcia, and M. A. Giorgetta (2006), Propagation of ENSO temperature signals into the middle atmosphere: A comparison of two general circulation models and ERA-40 reanalysis data, *J. Geophys. Res.*, *111*, D06101, doi:10.1029/2005JD006061.
Haynes, P. H., C. J. Marks, M. E. McIntyre, T. G. Shepherd, and K. P. Shine (1991), On the "downward control" of extratropical diabatic circulations by eddy-induced mean zonal forces, *J. Atmos. Sci.*, *48*, 651–678.
Held, I. M., and B. J. Soden (2000), Water vapor feedback and global warming, *Annu. Rev. Energy Environ.*, *25*, 441–475.
Held, I. M., and M. J. Suarez (1994), A proposal for the intercomparison of the dynamical cores of atmospheric general circulation models, *Bull. Am. Meteorol. Soc.*, *75*, 1825–1830.
L'Heureux, M. L., and D. W. J. Thompson (2006), Observed relationships between the El Niño–Southern Oscillation and the extratropical zonal-mean circulation, *J. Clim.*, *19*, 276–287, doi:10.1175/JCLI3617.1.
Li, F., J. Austin, and J. Wilson (2008), The strength of the Brewer-Dobson circulation in a changing climate: Coupled chemistry-climate model simulations., *J. Clim.*, *21*, 40–57.
Lin, P., Q. Fu, S. Solomon, and J. Wallace (2009), Temperature trend patterns in Southern Hemisphere high latitudes: Novel indicators of stratospheric change, *J. Clim.*, *22*, 6325–6341.
Lu, J., G. A. Vecchi, and T. Reichler (2007), Expansion of the Hadley cell under global warming, *Geophys. Res. Lett.*, *34*, L06805, doi:10.1029/2006GL028443.
Lu, J., G. Chen, and D. M. W. Frierson (2008), Response of the zonal mean atmospheric circulation to El Niño versus global warming, *J. Clim.*, *21*, 5942–5959.
McLandress, C., and T. G. Shepherd (2009), Simulated anthropogenic changes in the Brewer-Dobson circulation, including its extension to high latitudes, *J. Clim.*, *22*, 1516–1540.
Miller, R. L., G. A. Schmidt, and D. T. Shindell (2006), Forced annular variations in the 20th century Intergovernmental Panel on Climate Change Fourth Assessment Report models, *J. Geophys. Res.*, *111*, D18101, doi:10.1029/2005JD006323.
Neale, R. B., and B. J. Hoskins (2000), A standard test for AGCMs including their physical parameterizations. I: The proposal, *Atmos. Sci. Lett.*, *1*, 101–107.
Plumb, R. A. (2002), Stratospheric transport, *J. Meteorol. Soc. Jpn.*, *80*, 793–809.
Ring, M. J., and R. A. Plumb (2008), The response of a simplified GCM to axisymmetric forcings: Applicability of the fluctuation-dissipation theorem, *J. Atmos. Sci.*, *65*, 3880–3898.
Seager, R., N. Harnik, Y. Kushnir, W. Robinson, and J. Miller (2003), Mechanisms of hemispherically symmetric climate variability, *J. Clim.*, *16*, 2960–2978.
Son, S. W., and S. Lee (2005), The response of westerly jets to thermal driving in a primitive equation model, *J. Atmos. Sci.*, *62*, 3741–3757.
Thompson, D. W. J., J. M. Wallace, and G. C. Hegerl (2000), Annular modes in the extratropical circulation. Part II: Trends, *J. Clim.*, *13*, 1018–1036.
Yin, J. H. (2005), A consistent poleward shift of the storm tracks in simulations of 21st century climate, *Geophys. Res. Lett.*, *32*, L18701, doi:10.1029/2005GL023684.

G. Chen, Department of Earth and Atmospheric Sciences, Cornell University, 1115 Bradfield Hall, Ithaca, NY 14853, USA. (gc352@cornell.edu)
J. Lu, Center for Ocean-Land-Atmosphere Studies, 4041 Powder Mill Rd., Ste. 302, Calverton, MD 20705-3106, USA.
R. A. Plumb, Program in Atmospheres, Oceans and Climate, Massachusetts Institute of Technology, 77 Massachusetts Ave., Cambridge, MA 02139, USA.