This model calculates changes in air mass properties following their outflow (with or without cooling and/or windpeed and/or stable O isotope anomalies) over the sea, where evaporation adds vapour (with its own stable O composition that derives from the sea surface properties). The evaporation lowers the temperature (SST) the surface-water mixed layer (depth of which is set with the property "mld" - for temperature change we also account for insolation "Ins"), and the freshwater removal by evaporation also changes its salinity (SSS). Evaporation is calculated in daily increments, and the travel time over water is determined by means of the travel distance ("Le" for Israel and "Le/2" for Crete). We calculate all air- and water-mass properties and changes over a unit area of 1 m^2. For air properties at the end-point of the trajectory to Crete, we use $spac_C$=1, and for Haifa (Israel) we use tuning factor $spac_H$, as explained in the main text. These tuning factors are set once, fot ensure that - in control scenarios - the Crete and Haifa relative humidities and precipitation values match with observations. Then, we keep everything constant with respect to "spac". The "spac" factor in effect stands for the lateral (or vertical) spreading of air masses as they move away from the North Aegean. But more importantly, it allows tuning of the model to modern observations with one single parameter.

At the end of Le, or Le/2, the end-result air mass is uplifted to "height". "Height" could be used as a tuning factor, but it is preferred here to pick a reasonable number from the literature, and then keep it constant throughout the entire exercise. As the air cools, its maximum moisture content becomes lower (its relative humidity increases). When relative humidity reaches 1, the difference between the original moisture content of the air and the maximum possible content determines the amount that is precipitated out. The isotopic composition of the evaporated vapour is calculated using an equation for fractionation relative to the surface-water mixed layer value (δsw0), and air vapour O isotope-change is calculated using an isotope mass balance between original advected air to the onset, and the addedd component due to evaporation from the sea. Rain-out is isotopically in equilibrium with the vapour, and Rayleigh distillation is included. Snow-line altitudes are determined using the surface temperature and the constant assumed lapse rate, as are negative degree days at 2km altitude in Crete to get a sense for snow/ice preservation potential..

The model allows 0, 1, 3, 5 ten-day anomalies in winter to simulate polar/continental air outbreaks. The modern control is considered to be the 1-outbreak-period per year scenario. Thus, we retain one case (the 0 anomalies case) that is less severe than control, and two (3 and 5 anomalies cases) that are more severe than today.

A range of parameter values is then determined for years with 0, 1, 3, or 5 anomalies/events. Next, an approximation is made in which the GISP K+ timeseries is used to make a time-series of interval classes (0, 1, 3, 5 events per year based on the PDF of GISP K+ values). This is then used with the various year-type properties to develop timeseries that can be smoothed with different windows to simulate palaeo-proxy records with different levels of smoothing, which is what would be encountered in reality.

Finally, artificial conical lake settings are determined for Crete and the Levant (at sea level), in which the catchment area is fixed in a ratio relative to the lake surface so that the lakes are in steady state at control conditions. Lake-level variations are calculated assuming that all rainfall onto the catchment instantaneously enters the lake, and that evaporation applies only to the actual lake-water surface. The calculated lake-level changes illustrate trends to net aridity (drops) or humidity (rises), in a physically coherent manner to all the other parameters calculated by the model. A similar exercise is made using actual proportional relationships for the Dead Sea, to assess the model's implications for that basin. In that case, a limited amout of the precipitation over the catchment area is preserved (rest is lost to re-evaporation). Also evaporation rates from the lake itself are reduced by a factor α to account for the influence of the basin's high salinity (Rohling, 2013 QSR)

Due to specific MATHCAD quirks, the main model operator parameters/values are set (yellow highlights) on pages 9 and 32 .

For precipitation, we in addition work out the "amount effect", for comparison with observations close to Soreq Cave in Israel.

GISPK := t-K-K200.xls

$GISPt := GISPK^{\langle 0 \rangle}$ $GISPKraw := GISPK^{\langle 1 \rangle}$ $GISPK200 := GISPK^{\langle 2 \rangle}$

0

$yeardays := 365$ $yearN := 4$

$i := 0 .. yearN \, yeardays - 1$

Total pathway over eMed from N Aegean to Israel is roughly 1400 km
For N Aegean to Crete we use half that.

$LeH := 1400 \cdot 10^3 \, m$

$LeC := \frac{LeH}{2}$

Incoming air flow

Wind speed (double peak annually in mid winter and in mid summer) after Santorini: data on
https://weatherspark.com/averages/32207/Santorini-Thira-South-Aegean-Greece

mean incoming air Temp $Ta0mean := 12$

seasonal inc air T range $dTa0 := 22$

Based on approx **Plovdiv, Bulgaria, mean monthly T range from 1 to 23 deg C, with annual mean of 12.**
http://www.plovdiv.climatemps.com

Relative humidity

mean incoming relative humidity $r0mean := 0.6$

seasonal inc rel humidity range $dr0 := 0.3$

set seasonal cycle for incoming rel. humidity:

$$r0_i := r0mean - \frac{dr0}{2} \cdot \sin\left(\frac{i \cdot 2\pi}{yeardays}\right)$$

Plovdiv, Bulgaria annual range is 0.45-0.75

(Alexandropolis, Greece range is 0.55-0.75 acc to climatemps

We assume that the relative humidity changes (increases) with height, linearly in proportion to reduction (lapse rate) of Ta. Doing so means that we can calculate total average vapour load based on surface lapse rate, as specific humidity is kept constant but relative humidity changes.

Insolation

$Imean := 310$ $dI := 300$

mean incoming insolation at top of atmosphere (TOA)

$$I_i := Imean + \frac{dI}{2} \cdot \sin\left(\frac{i \cdot 2\pi}{yeardays}\right) \quad \frac{W}{m^2}$$

seasonal inc insolation range

$$a_i := 0.5 + \frac{0.3}{2} \cdot \sin\left(\frac{i \cdot 2\pi}{yeardays}\right)$$

$$Ins_i := a_i \cdot I_i$$

set a (=1-α) to go from 0.25 in winter to 0.55 in summer, so as to get **received surface insolation** as in Athens observations. Normal places are cleaner than Athens, so we use a range of 0.35 - 0.65

For Athens, received irradiation goes from about 50 to 225 W/m^2, acc. to http://www.mdpi.com/2071-1050/6/8/5354/htm

Initial ocean condition

mean Sea Surface Temp $Ts0mean := 18$

seasonal SST range $dTs0 := 10$

T range from Rohling et al (2002 Clim Dyn), as 13 to 23 deg C.

$$Ts0_i := Ts0mean + \frac{dTs0}{2} \cdot \sin\left(\frac{i \cdot 2\pi}{yeardays}\right)$$

Air layer depth

$$D_i := \left(1200 + \frac{400}{2} \cdot -\sin\left(\frac{i \cdot 2\pi}{yeardays}\right)\right)$$

We set D to cycle between about 1000 and 1400 m (see main text).

Depths and phasing according to Dayan et al. (2017 Atmos Chem Phys)

Mean landfall (Crete/Heraklion) air Temp (at sea level)

$TaCmean := 22$

seasonal inc air T range $dTaC := 14$

Santorini T range is 12-27 degrees.
http://www.climatedata.eu/climate.php?loc=grxx0030&lang=en
Heraklion T range is 15-29 degrees.

mean landfall (Haifa) air Temp (at sea level)

$Tacmean := 21$

Haifa range is 14-28 deg C according to climatemps

seasonal inc air T range $dTac := 14$

We now import a file in which the same sinusoidal background signals (as outlined above) are perturbed with 1, 3, or 5 ten-day events in winter. The file starts with age, and then lists TaO (for events between -0 and -20 deg C), V (with events between 0 and 20 m/s), TaC0 for Crete (with events between -0 and -20 deg C), TaH0 for Haifa (with events between -0 and -20 deg C), and δa0 the isotopic composition of incoming air vapour (with events between 0 and -2 permil).

$M :=$ ıge_Ta0_V_TaC0_TaH0_da0.xls $\qquad age := M^{\langle 0 \rangle} \qquad length(age) = 1460$

Read out the various parameter files from "M"

$Ta0m := submatrix(M, 0, 1459, 1, 21) \qquad TaC0m := submatrix(M, 0, 1459, 43, 63) \qquad TaH0m := submatrix(M, 0, 1459, 64, 84)$

$Vm := submatrix(M, 0, 1459, 22, 42) \qquad \delta a0m := submatrix(M, 0, 1459, 85, 105)$

Read out the records according to different selections for the perturbation intensities

$Ta0 := Ta0m^{\langle n_T \rangle}$ $V0 := Vm^{\langle n_V \rangle}$ $TaC0 := TaC0m^{\langle n_T \rangle}$ $TaH0 := TaH0m^{\langle n_T \rangle}$ $\delta a0 := \delta a0m^{\langle n_i \rangle}$

Wind attenuation over distance due to air column spreading ("spacH"):

Then, mean wind speeds over distances to Crete and Haifa are:

$VC_i := V0_i$ $VH_i := \dfrac{VC_i}{spacH}$ $VCm := \dfrac{V0 + VC}{2}$ $VHm := \dfrac{VC + VH}{2}$

Set height so it is lifting air 0 m in summer and 2xhme m in winter (to simulate summer air stability/troposheric descent, which suppresses summer rainfall; see main text).

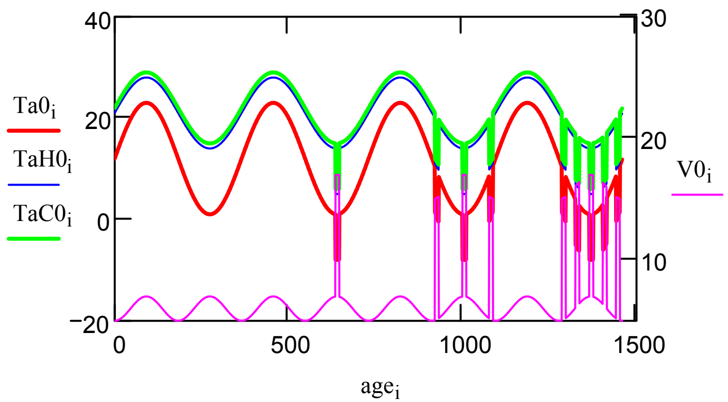
$hme \equiv 610$ $he_i := hme - \dfrac{2 \cdot hme}{2} \cdot \sin\left(\dfrac{i \cdot 2\pi}{yeardays}\right)$ $\min(he) = 0$ $\max(he) = 1220$

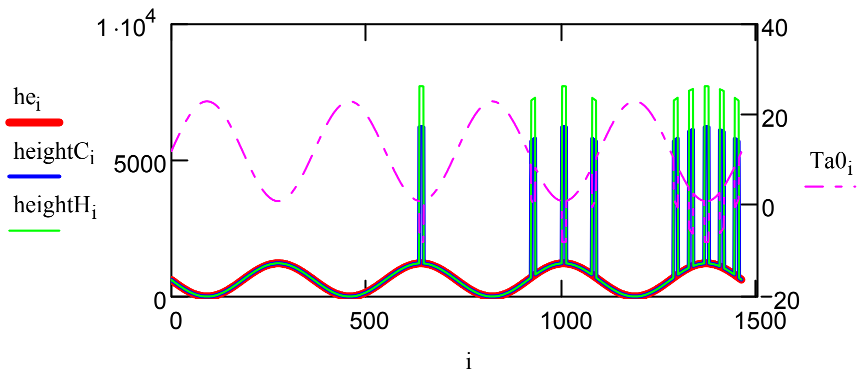
hme is used to tune the model to give correct rainfall values for Crete, returned below

set selector for "height" so that the value "maxhei" is returned for the cold events. Because of rounding, the two critical terms are not exactly equal, so need to set a threshold of one tenthousandth (maxheiC and maxheiH are set below on the settings-page, page 9).

$$heightC_i := \begin{cases} maxheiC + he_i & \text{if } Ta0_i < Ta0mean + \dfrac{dTa0}{2} \cdot \sin\left(\dfrac{i \cdot 2\pi}{yeardays}\right) - 0.0001 \\ he_i & \text{otherwise} \end{cases}$$

$$heightH_i := \begin{cases} maxheiH + he_i & \text{if } Ta0_i < Ta0mean + \dfrac{dTa0}{2} \cdot \sin\left(\dfrac{i \cdot 2\pi}{yeardays}\right) - 0.0001 \\ he_i & \text{otherwise} \end{cases}$$





In the model, we use a year with one event as the modern "control" year. That way, we can have years that are more benign (no events), and years that are more severe (i.e., 3 or 5 events).

Next calculate air-mass moisture changes

Start with those over the Aegean sector (Northern Greece to Crete)

Exchange coefficient: $C := 1.15 \cdot 10^{-3}$ (Garrett et al., 1993; Wells, 1995)

Density of air: $\rho \equiv 1.27$ kg m^{-3} (at mean sea level pressure = 1013 mbar)

Total pressure $p \equiv 1.013 \cdot 10^{5}$ Pa at mean sea level

Values from page 41 (citing Korres et al., 2000) of https://books.google.com.au/books?id=w0bVBgAAQBAJ&pg=PA34&lpg=PA34&dq=aegean+and+albedo&source=bl&ots=QAGgvMZ-FQ&sig=65GJ-8H1L-tAa0e9RK4AF0IAur0&hl=en&sa=X&ei=XBEnVe7WOM6F8gWN-YDQDA&ved=0CE4Q6AEwCA#v=onepage&q=aegean%20and%20albedo&f=false

Following Abbott and Tabony (1985):
rewritten to solve for T in Celcius and to give results in Pa:

$$L_i := \left(2500.84 - 2.34 \cdot Ts0_i\right) \cdot 10^{3}$$ J kg^{-1}

Use lapse rate to calculate air T change due to uplift over Crete

$$\Delta TvC_i := \frac{heightC_i}{1000} \cdot LR$$

over Haifa (Israel)

$$\Delta TvH_i := \frac{heightH_i}{1000} \cdot LR$$

and to the mean height of air column D:

$$TaD0_i := Ta0_i - \frac{0.5\,D_i \cdot LR}{1000}$$

Calculate saturation equation at sea surface temperature

$$es_i := 10^{2} \cdot e^{55.17 - 6803 \cdot \left(Ts0_i + 273.15\right)^{-1} - 5.07 \cdot \ln\left(Ts0_i + 273.15\right)}$$

Calculate saturation equation at air temperature (mean of D) during evaporation

$$ea_i := 10^{2} \cdot e^{55.17 - 6803 \cdot \left(TaD0_i + 273.15\right)^{-1} - 5.07 \cdot \ln\left(TaD0_i + 273.15\right)}$$

Then, following Wells (1986; p.83) $qs_i := \frac{es_i}{p - es_i} \cdot \left(\frac{18.0153}{28.965}\right)$

$$qa_i := \frac{ea_i}{p - ea_i} \cdot \left(\frac{18.0153}{28.965}\right)$$

Calculate evaporation (see Wells and King-Hele, 1990):

$$EC_i := -\rho \cdot L_i \cdot C \cdot VCm_i \cdot \left(qs_i - r0_i \cdot qa_i\right) \cdot 1.26 \cdot 10^{-2}$$ m yr^{-1}

based on 1 W equivalent to 1.26 cm yr^{-1} = 1.26*10^{-2} m yr^{-1} per unit area (m^{-2}).(Wells, 1986, p.127; Garrett et al., 1993)

$$LHC_i := -\rho \cdot L_i \cdot C \cdot VCm_i \cdot \left(qs_i - r0_i \cdot qa_i\right)$$ Watt (= J/s) per m^2

Where the latter equation (LH) gives the Latent Heat

Do a scaling check on the daily and annual T loss due to evaporation, versus T gain from insolation

specific heat capacity of water: $SHC := 4.187 \cdot 10^3 \frac{J}{kg \cdot K}$

See http://www.jgsee.kmutt.ac.th/exell/JEE661/JEE661Lecture2.html

no. of seconds in a day: $\mu := 60 \cdot 60 \cdot 24$ $\mu = 8.64 \times 10^4$

For SST & SSS changes, we here choose to specifically calculate these for the Aegean (down to Crete), where we have observations.

LH*m gives the daily heat loss in Joules/m^2

We are looking at a loss per m^2 from a surface layer of "mld" thickness, i.e., from mld m^3 of water, which is about 1.040*mld*10^3 kg

$$\Delta heat\_evap_i := \frac{(LHC_i) \cdot \mu}{SHC} \quad \frac{kg \cdot K}{m^2} \qquad \Delta heat\_tot_i := \frac{(Ins_i + LHC_i) \cdot \mu}{SHC}$$

Daily change in surface water T due to evaporative heat loss:

$$\Delta Tsevap_i := \frac{\Delta heat\_evap_i}{1.040 \cdot mld \cdot 1000} \qquad \Delta Tstot_i := \frac{\Delta heat\_tot_i}{1.040 \cdot mld \cdot 1000}$$

where "mld" is the mixed layer depth set in the starting conditions

First we look at the annual T loss due to evaporation in the 'control' years (i.e., 1-event years):

$$ann\_evapchange := \sum_{i=365}^{2 \cdot 365 - 1} \Delta Tsevap_i \qquad ann\_evapchange = -15.001$$

In measurements, seasonal SST range is about 10 degrees (13 to 23 deg C -see above) - and our Evaporation-caused T loss must be approx. similar

Next, the total annual T change (after insolation warming is added) is:

$$annchange := \sum_{i=365}^{2 \cdot 365 - 1} \Delta Tstot_i \qquad annchange = -2.961$$

So the annual T loss from an "mld" m layer of water is given, just from evaporative cooling. The observed Aegean range of 10 deg C is closely found when mld = 100. Clearly, evaporative heat loss explains most of the seasonal cycle of T in the surface waters of the Aegean. SST cycles back up due to irradiation. So we can satisfactorily view the seasonal cycle this way, and only calculate the SST change **anomalies** associated with perturbations to the regular seasonal evaporation cycle, due to imposed polar air outbreaks.

By bringing in realistic insolation values, the annual net change is small - just as it should be (seeing that we are not accounting for conductive heat loss, or heat advection). Clearly, the vast bulk of the energy for evaporation comes from the insolation (as it should be).

In the "event" years, we find:

$$ann\_evapchange\_0event := \sum_{i=0}^{364} \Delta Tsevap_i \qquad annchange0event := \sum_{i=0}^{364} \Delta Tstot_i$$

$$ann\_evapchange\_0event = -14.283 \qquad annchange0event = -2.243$$

We can also calculate the Aegean annual mean salinity anomalies:

$$dSSS_i := \frac{SSS1 \cdot mld}{mld + \frac{EC_i}{365}} - SSS1$$

$SSS1 := 39$

$$\Delta SSS0 := \sum_{i=0}^{364} dSSS_i$$

$\Delta SSS0 = 0.9692$

$$ann\_evapchange\_3event := \sum_{i=2\cdot 365}^{(3\cdot 365)-1} \Delta Tsevap_i$$

$$ann\_evapchange\_3event = -16.754$$

$$annchange3event := \sum_{i=2\cdot 365}^{(3\cdot 365)-1} \Delta Tstot_i$$

$$annchange3event = -4.714$$

$$\Delta SSS1 := \sum_{i=365}^{(2\cdot 365)-1} dSSS_i$$

$$\Delta SSS1 = 1.018$$

$$\Delta SSS3 := \sum_{i=2\cdot 365}^{(3\cdot 365)-1} dSSS_i$$

$$\Delta SSS3 = 1.1369$$

$$ann\_evapchange\_5event := \sum_{i=3\cdot 365}^{(4\cdot 365)-1} \Delta Tsevap_i$$

$$ann\_evapchange\_5event = -18.275$$

$$annchange5event := \sum_{i=3\cdot 365}^{(4\cdot 365)-1} \Delta Tstot_i$$

$$annchange5event = -6.235$$

$$\Delta SSS5 := \sum_{i=3\cdot 365}^{(4\cdot 365)-1} dSSS_i$$

$$\Delta SSS5 = 1.2401$$

So we find annual SST change anomalies of the order of:

$annchange0event - annchange = 0.719$ degrees C

$annchange3event - annchange = -1.753$

$annchange5event - annchange = -3.273$

$\Delta SSS0 - \Delta SSS1 = -0.049$ psu

$\Delta SSS3 - \Delta SSS1 = 0.119$

$\Delta SSS5 - \Delta SSS1 = 0.222$

Bear in mind that, in SST, the events in proxy data seem to be of the order of 1 to 2 deg C, so we would want to see approx agreement here.

Calculate the moisture addition to the incoming air (added to the initial moisture load). To do so, the distance over which evaporation adds moisture is Le for the entire distance to Haifa, and half that (Le/2) for the distance to Crete.

We are adding flux E (in daily proportions, so E/365) into air column of depth D, temperature Ta0, and relative humidity r. Consequently, r moves toward saturation.

Travel time over water: Crete = $TtC_i := \frac{LeC}{VCm_i}\cdot\frac{1}{\mu}$ days Haifa = $TtH_i := \frac{LeH}{VHm_i}\cdot\frac{1}{\mu}$ days The 1/μ term scales things up to measures per day (m is 86400 seconds in a day).

$$dayEadditionC_i := \frac{(-EC)_i}{365}$$ in m/day

addition to the air, by weight (per unit area)

$dayEadwC_i := dayEadditionC_i\cdot 1000$ in kg of freshwater per m^2

r0*qa is expressed in (10^-3) kg water per kg of air So in a colum of thickness D, there is a mass of moisture per m^2 as determined D*r0*qa/Dmean(weight of air per m^3):

Density of air at sea level = ρ $\rho = 1.27$ kg m^{-3} (at mean sea level pressure = 1013 mbar)

Work out mean pressure and density for D

$$pmean_i := p \cdot e^{\frac{0.5 \cdot D_i}{7400}} \quad \text{Pa} \qquad \rho mean_i := \rho \cdot e^{\frac{0.5 \cdot D_i}{7400}} \quad \frac{kg}{m^3}$$

$$p(z) = p0 \cdot e^{\frac{z}{H}} \qquad \rho(z) = \rho 0 \cdot e^{\frac{z}{H}}$$ where H=7.4 km

http://acmg.seas.harvard.edu/people/faculty/djj/book/bookchap2.html

Calculate saturation state for Crete/Santorini Temps

$$TaDC0_i := TaC0_i - \frac{0.5\, D_i \cdot LR}{1000}$$

$$eaC_i := 10^2 \cdot e^{55.17 - 6803 \cdot (TaDC0_i + 273.15)^{-1} - 5.07 \cdot \ln(TaDC0_i + 273.15)}$$

$$qaC_i := \frac{eaC_i}{pmean_i - eaC_i} \cdot \left(\frac{18.0153}{28.965}\right)$$

Calculate saturation state for Haifa Temps

$$TaDH0_i := TaH0_i - \frac{0.5\, D_i \cdot LR}{1000}$$

These are needed to calculate saturation pressures at mean Temp of column D

$$eaH_i := 10^2 \cdot e^{55.17 - 6803 \cdot (TaDH0_i + 273.15)^{-1} - 5.07 \cdot \ln(TaDH0_i + 273.15)}$$

$$qaH_i := \frac{eaH_i}{pmean_i - eaH_i} \cdot \left(\frac{18.0153}{28.965}\right)$$

Because we assume specific humidity is constant through D, initial moisture contained in D is:

$$M0_i := \rho mean_i \cdot r0_i \cdot qa_i \cdot D_i \quad \frac{kg}{m^2}$$

and Max moisture contained at the end-points (C and H) is:

$$MmaxC_i := \rho mean_i \cdot 1 \cdot qaC_i \cdot D_i \qquad MmaxH_i := \rho mean_i \cdot 1 \cdot qaH_i \cdot D_i$$

Now add the daily evap over Tt timesteps, and then can calculate the saturation state again.
We will assume that up to Tt*dayEadw is added, OR **less** until r endvalue (rend) =1

This is at the surface, at Ta0. At higher levels T is lower and so rend (as r) is higher, even if air and vapour are well mixed, because of lapse rate, and because we assume constant specific humidity in the column of thickness D!

max moisture potential for Crete air

$$MendC\_a_i := MmaxC_i$$

Actual moisture at <u>end of passage</u>

$$MendC\_b_i := \left(M0_i \cdot \frac{\mu\, VCm_i}{LeC} + dayEadwC_i\right) \cdot TtC_i$$

Note that traveltime is in DAYS!

$$MendC_i := \begin{cases} MendC\_b_i & \text{if } MendC\_b_i < MendC\_a_i \\ MendC\_a_i & \text{otherwise} \end{cases}$$

Relative humidity of the air (at Temp of meanD) at Crete:

$$rendC_i := \frac{MendC_i}{MmaxC_i}$$

Work out precip when Ta drops due to uplift of the air by "heightC" for Crete, or "heightH" for Haifa

$pmeanupC_i := p \cdot e^{\frac{0.5 D_i + heightC_i}{7400}}$ $\quad \rho meanupC_i := \rho \cdot e^{\frac{0.5 D_i + heightC_i}{7400}}$ $\quad pmeanupH_i := p \cdot e^{\frac{0.5 D_i + heightH_i}{7400}}$ $\quad \rho meanupH_i := \rho \cdot e^{\frac{0.5 D_i + heightH_i}{7400}}$

For Crete: at condensation air temperature TaDC0-ΔTv deg C

$$eacC_i := 10^2 \cdot e^{55.17 - 6803 \cdot (TaDC0_i - \Delta TvC_i + 273.15)^{-1} - 5.07 \cdot \ln(TaDC0_i - \Delta TvC_i + 273.15)}$$

$$qacC_i := \frac{eacC_i}{pmeanupC_i - eacC_i} \cdot \left(\frac{18.0153}{28.965}\right)$$

Now we have moisture load MendC coming in, and the max. moisture load M1C that might be held in uplifted air at relH of 1.
The difference between MendC and M1C (when POSITIVE) is a surplus of water that needs to rain out.

$$M1C_i := \rho meanupC_i \cdot 1 \cdot qacC_i \cdot D_i \qquad \frac{kg}{m^2}$$

Fraction of remaining atm. vapour remaining: $FC_i := \frac{M1C_i}{MendC_i}$ $\quad fC_i := \begin{cases} FC_i & \text{if } FC_i \le 1 \\ 1 & \text{otherwise} \end{cases}$

MmaxCout (max moisture load carried in air leaving Crete) at T after dry adiabatic descent (dry adiabatic lapse rate almost 10 C/km)

$$TaDCout_i := TaDC0_i - \Delta TvC_i + heightC_i \cdot \frac{10}{1000}$$

$$eaCout_i := 10^2 \cdot e^{55.17 - 6803 \cdot (TaDCout_i + 273.15)^{-1} - 5.07 \cdot \ln(TaDCout_i + 273.15)}$$

MmaxCout is the max moisture load at T after dry adiabatic descent

$$qaCout_i := \frac{eaCout_i}{pmean_i - eaCout_i} \cdot \left(\frac{18.0153}{28.965}\right)$$

$$MmaxCout_i := \rho mean_i \cdot 1 \cdot qaCout_i \cdot D_i \qquad rCout_i := \frac{fC_i \cdot MendC_i}{MmaxCout_i}$$

Now repeat all calculations for Haifa, using air that "evolved" over Crete

evaporation: $EH_i := -\rho \cdot L_i \cdot C \cdot VHm_i \cdot (qs_i - rCout_i \cdot qaC_i) \cdot 1.26 \cdot 10^{-2}$ $\quad LHH_i := -\rho \cdot L_i \cdot C \cdot VHm_i \cdot (qs_i - rCout_i \cdot qaC_i)$

$dayEadditionH_i := \frac{(-EH)_i}{365}$ in m/day $\qquad dayEadwH_i := dayEadditionH_i \cdot 1000$ in kg of freshwater per m^2

max moisture potential for Haifa air

$MendH\_a_i := MmaxH_i$

Actual moisture at end of passage: $MendH\_b_i := \frac{\left(MendC_i \cdot fC_i \cdot \frac{\mu\, VHm_i}{LeH - LeC} + dayEadwH_i\right) \cdot (TtH_i - TtC_i)}{spacH}$

$$MendH_i := \begin{cases} MendH\_b_i & \text{if } MendH\_b_i < MendH\_a_i \\ MendH\_a_i & \text{otherwise} \end{cases}$$

rel. hum of air arriving at surface Haifa (rendH): $rendH_i := \frac{MendH_i}{MmaxH_i}$

For "spacH", see parameter settings page (9)

continued on page 10

SETTINGS PAGE

Repeat values (set on page 32) for the adjustable parameters for experiments:

mixed layer depth $mld = 100$

$Vinc := n_V \cdot 1$ $Vinc = 10$ $\max(V) = 1\,V$

Vinc up to 10 would even be reasonable as Bora winds can be consistent at 17 m/s
http://www.istrianet.org/istria/geosciences/meteorology/winds/bora-adriatic.htm

$cool := n_T \cdot 1$ $cool = 9$ $\min(TaC0) = 6$

$\min(TaH0) = 5$

Cooling may be set up to 9-10 deg C for an event as the wind entering is feezing (Casford et al., 2003). Below we argue why 6 seems the most representaitve, given the isotope results versus isotope data.

$\Delta\delta a := n_i \cdot -0.1$ $\Delta\delta a = 0$

Setting (once!) values for model tuning:

$maxheiC \equiv 5000$

$maxheiH \equiv 6500$

Ben Ami et al. (2015 Nat. Haz. Earth Syst. Sci.) did a thunderstrom study and found that convective layer thickness in least stable conditions was thinnest, at around 5 km, so set that value. Also Galvin et al (2011 Weather) seem to suggest that 5 km is representative. Same value from Altaratz et al. (2001 Monthly Weather Rev).
Ganot et al. (2007 GRL) reports observations for Cyprus Low systems reaching at least 6.5 km

$mld \equiv 100$ sea mixed layer depth

$LR \equiv 6.5$ Lapse rate over Israel typically between **6.5** and 7 (see main text)

$spacH \equiv 1.595$

SPATIAL expansion of column D, via an 'apparent' height calculation with a multiplier
This means we are adding 1/spac times the moisture to our column D through evaporation

Need to tune this manually (once!)

"spacH" is used for tuning to give annual P values similar to the observed annual P values, for control years.
The value for Crete is kept to 1, through tuning with "hme" above. That is then kept constant, and Haifa values are tuned using spacH.
We use spacH as a tuning factor for the control years, and then keep it constant while doing scenarios that look at relative changes in the event years. Note that wind speeds are proportionally changed (in a linear manner) to ensure conservation of mass.

$\delta sw0 \equiv 1.5$ Initial surface sea-water δsw: should be close to 1.5 permil, after https://data.giss.nasa.gov/o18data/

continued from p.8

$$eacH_i := 10^2 \cdot e^{55.17 - 6803 \cdot (TaDH0_i - \Delta TvH_i + 273.15)^{-1} - 5.07 \cdot \ln(TaDH0_i - \Delta TvH_i + 273.15)}$$

$$qacH_i := \frac{eacH_i}{pmeanupH_i - eacH_i} \cdot \left(\frac{18.0153}{28.965}\right)$$

Now we have moisture load MendH coming in, and the max. moisture load M1H that <u>might</u> be held in uplifted air at relH of 1.
The difference between MendH and M1H (when POSITIVE) is a surplus of water that needs to rain out.

$$M1H_i := \rho meanupH_i \cdot 1 \cdot qacH_i \cdot D_i \quad \frac{kg}{m^2}$$

Fraction of remaining atm. vapour: $FH_i := \dfrac{M1H_i}{MendH_i}$

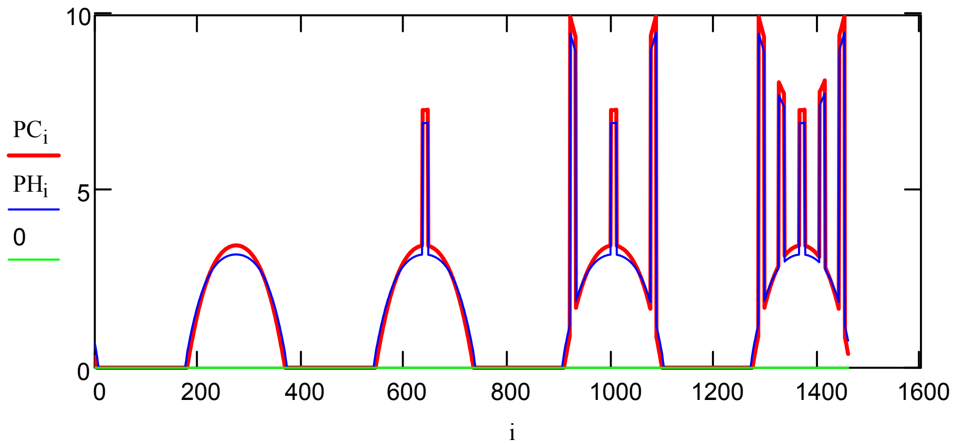
$$fH_i := \begin{cases} FH_i & \text{if } FH_i \le 1 \\ 1 & \text{otherwise} \end{cases}$$

The difference between Mend and M1 both for C and for H is the amount of precipitation that has to follow uplift, in kg/m^2. Thus we can calculate P in mm for each region by considering that weight in mm of P over a sq m.

1 kg/m^2 = 1 mm/m^2

Precip amounts (in mm per m^2) then are:

$$PC_i := \begin{cases} MendC_i - M1C_i & \text{if } MendC_i > M1C_i \\ 0 & \text{otherwise} \end{cases} \qquad PH_i := \begin{cases} MendH_i - M1H_i & \text{if } MendH_i > M1H_i \\ 0 & \text{otherwise} \end{cases}$$



Annual rainfall for Heraklion (PC) is **500** mm (see link above with T data) (of which 70 in summer)

Annual rainfall (PH) for Haifa **500** mm (Climatemps)
Jerusalem (measured) is 590 mm (amost 0 in summer)

Crete
Annual P
control:

$$\sum_{i=365}^{2 \cdot 365 - 1} PC_i = 504.918$$

Haifa
Annual P
control:

$$\sum_{i=365}^{2 \cdot 365 - 1} PH_i = 504.99$$

The blue highlighted numbers are tuned with hme and spacH - <u>we set things up so we get a reasonable match for the 1-anomaly case</u>

To determine E-P in Crete ***at sea level*** (e.g., Heraklion), we also need Evaporative loss there (in mm/day). **The loops below determine a matrix of Evap values at 100 ALTITUDE increments up to 1000 m altitude, so we could use results for higher locations too**:

For Crete:

$$
\begin{aligned}
\mathrm{EmmdayC} := \; & \text{for } i \in 0 .. \mathrm{yearN}\,\mathrm{yeardays} - 1 \\
& \quad \text{for } k \in 0 .. 10 \\
& \quad\quad L_{i,k} \leftarrow \left[2500.84 - 2.34 \cdot \left(\mathrm{TaC0}_i - \frac{k \cdot 100}{1000} \cdot \mathrm{LR}\right)\right] \cdot 10^3 \\
& \quad\quad \mathrm{eac}_{i,k} \leftarrow 10^2 \cdot e^{55.17 - 6803 \cdot \left[\left(\mathrm{TaC0}_i - \frac{k \cdot 100}{1000} \cdot \mathrm{LR}\right) + 273.15\right]^{-1} - 5.07 \cdot \ln\left[\left(\mathrm{TaC0}_i - \frac{k \cdot 100}{1000} \cdot \mathrm{LR}\right) + 273.15\right]} \\
& \quad\quad \mathrm{qac}_{i,k} \leftarrow \frac{\mathrm{eac}_{i,k}}{p - \mathrm{eac}_{i,k}} \cdot \left(\frac{18.0153}{28.965}\right) \\
& \quad\quad \mathrm{Emmday}_{i,k} \leftarrow \frac{-\rho \cdot L_{i,k} \cdot C \cdot \mathrm{VCm}_i \cdot \left(\mathrm{qac}_{i,k} - \mathrm{rendC}_i \cdot \mathrm{qac}_{i,k}\right) \cdot 1.26 \cdot 10^{-2}}{365} \cdot 1000 \\
& \mathrm{Emmday}
\end{aligned}
$$

For Haifa:

$$
\begin{aligned}
\mathrm{EmmdayH} := \; & \text{for } i \in 0 .. \mathrm{yearN}\,\mathrm{yeardays} - 1 \\
& \quad \text{for } k \in 0 .. 10 \\
& \quad\quad L_{i,k} \leftarrow \left[2500.84 - 2.34 \cdot \left(\mathrm{TaH0}_i - \frac{k \cdot 100}{1000} \cdot \mathrm{LR}\right)\right] \cdot 10^3 \\
& \quad\quad \mathrm{eac}_{i,k} \leftarrow 10^2 \cdot e^{55.17 - 6803 \cdot \left[\left(\mathrm{TaH0}_i - \frac{k \cdot 100}{1000} \cdot \mathrm{LR}\right) + 273.15\right]^{-1} - 5.07 \cdot \ln\left[\left(\mathrm{TaH0}_i - \frac{k \cdot 100}{1000} \cdot \mathrm{LR}\right) + 273.15\right]} \\
& \quad\quad \mathrm{qac}_{i,k} \leftarrow \frac{\mathrm{eac}_{i,k}}{p - \mathrm{eac}_{i,k}} \cdot \left(\frac{18.0153}{28.965}\right) \\
& \quad\quad \mathrm{Emmday}_{i,k} \leftarrow \frac{-\rho \cdot L_{i,k} \cdot C \cdot \mathrm{VHm}_i \cdot \left(\mathrm{qac}_{i,k} - \mathrm{rendH}_i \cdot \mathrm{qac}_{i,k}\right) \cdot 1.26 \cdot 10^{-2}}{365} \cdot 1000 \\
& \mathrm{Emmday}
\end{aligned}
$$

For altitudes of 0, 100, 200 ...1000 m above Heraklion and Haifa, we can now determine the E-P balance: $k := 0..10$

For Crete: $netFC_{i,k} := EmmdayC_{i,k} + PC_i$ in mm

For Levant: $netFH_{i,k} := EmmdayH_{i,k} + PH_i$ in mm

Net E-P balances (in mm) at sea level (k=0)

$$\sum_{i=365}^{2\cdot 365-1} \left(netFC^{\langle 0\rangle}\right)_i = -526$$
Control YEAR

$$\sum_{i=365}^{365+183-1} \left(netFC^{\langle 0\rangle}\right)_i = -722$$
Control SUMMER

$$\sum_{i=365+183}^{(2\cdot 365)-1} \left(netFC^{\langle 0\rangle}\right)_i = 196$$
Control WINTER

$$\sum_{i=0}^{364} \left(netFC^{\langle 0\rangle}\right)_i = -587$$
0 Event YEAR

$$\sum_{i=0}^{182} \left(netFC^{\langle 0\rangle}\right)_i = -722$$
0 Event SUMMER

$$\sum_{i=183}^{364} \left(netFC^{\langle 0\rangle}\right)_i = 135$$
0 Event WINTER

$$\sum_{i=2\cdot 365}^{(3\cdot 365)-1} \left(netFC^{\langle 0\rangle}\right)_i = -298$$
3 Event YEAR

$$\sum_{i=2\cdot 365}^{2\cdot 365+183-1} \left(netFC^{\langle 0\rangle}\right)_i = -722$$
3 Event SUMMER

$$\sum_{i=2\cdot 365+183}^{(3\cdot 365)-1} \left(netFC^{\langle 0\rangle}\right)_i = 424$$
3 Event WINTER

$$\sum_{i=3\cdot 365}^{(4\cdot 365)-1} \left(netFC^{\langle 0\rangle}\right)_i = -152$$
5 Event YEAR

$$\sum_{i=3\cdot 365}^{3\cdot 365+183-1} \left(netFC^{\langle 0\rangle}\right)_i = -722$$
5 Event SUMMER

$$\sum_{i=3\cdot 365+183}^{(4\cdot 365)-1} \left(netFC^{\langle 0\rangle}\right)_i = 570$$
5 Event WINTER

Because all events are in winter, the summer values don't change between different years.

Note that all values are in mm (not in mm/y), so annual values need to be obtained by addition of summer and winter values.

Now use Lapse Rates and surface temperatures to approximate snow-line altitudes in Crete and Haifa/Levant. This is important because key evidence for the RCC events consists of Little Ice Age reports of snow and frost. Also today, outbreak events are characterised by snow in unusual places (Athens, Istanbul).

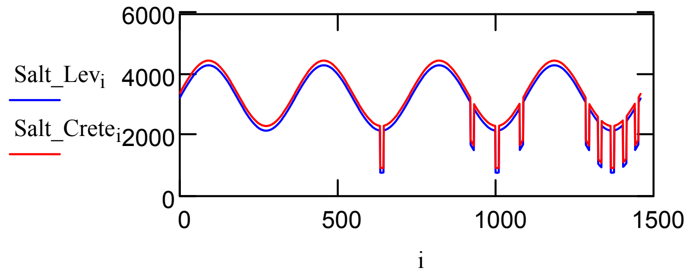
Cretan snowline altitude $Salt\_Crete_i := \frac{TaC0_i}{LR} \cdot 1000$

Levant snowline altitude $Salt\_Lev_i := \frac{TaH0_i}{LR} \cdot 1000$

Lowest altitude of snowline during events

$min(Salt\_Crete) = 923$

$min(Salt\_Lev) = 769$



Xoplaki et al (2001 Clim. Cha.): "*From the scattered data found for 1675-1715 and 1780-1830, the winter and spring climate in southern Balkans and the eastern Mediterranean, especially during the LMM, can be characterised as cooler and relatively rainier with a higher variability compared with the recent decades.*"

Event cooling factor "cool" is set on the Key operator values page (32)

Snowline today in the western Taurus Mountains, Turkey is at about 3400-3700 m.

Cretan mountains reach up to almost 2500 m. Typically, Crete's three key mountain ranges reach above 2000 m.
Event periods had persistent snow on the ranges, so snowline altitude 2000 m or lower.

At "cool" = 0, lowest snowline is at 2460 m. At "cool" = 1 it is at 2315 m. At "cool" = 2 it is at 2152 m. At "cool" = 3 it is at 1990 m. At "cool" = 4 it is at 1826 m. At "cool" = 5 it is at 1662 m. At "cool" = 6 it is at 1500 m. So **using both snowline and SST, we estimate "cool" > 3 C.**

Today, Thessaloniki air T range through year: 5 -26.5 deg C.
So to get <u>long-lasting frosts in winter in Thessaloniki</u> (at sea level in northernmost Aegean), we need "cool" > 5 deg C.

Xoplaki et al. (2001) write: In Thessaloniki, the winter of 1954 was the coldest for the 1901-1998 period. Thessaloniki reported 25 snowy days from December 1953 to February 1954 (mean of the period 1950-1998 is 8 snowfall events; Mparsakis, 1999) and two times the normal amount of precipitation compared to the long-term average (1901-1998) of 120 mm. In addition, for Athens below normal temperatures and above normal precipitation values were measured

Note: Most extreme events:

If snow is to fall and hold (for the duration of the event) in Athens, Heraklion, etc. (coastal cities in S Aegean), then we need "cool" to be ~15 degrees C. Winter temps in Thessaloniki would then plummet to around -10 deg C. P events would be extreme, causing much flash-flooding (NB. that is true already at "cool" = 5). The NOAA weather log for 1-10 december 2001 (below) suggests that such events do happen!
If sustained/repeated in winter, this would give about 2.6 deg C SST change in winter relative to Control - again this is NOT impossible.

Good middle way uses event-mean "cool" = 9 C, understanding that, for severe extremes, "cool" may (briefly) reach 15 deg C. For "cool" = 9, the Levant amount effect in δ18O is -0.5 permil/100 mm of rainfall, as observed (see main text).

ftp://ftp.ncdc.noaa.gov/pub/data/extremeevents/specialreports/Climate-Watch-December-2001.pdf

Weather Log - December 1-10th, 2001
According to media reports **coastal flooding in Turkey has left at least three people dead. Hundreds of homes and businesses have been drenched by five consecutive days of rain, and flood waters have swept away roads, a bridge and a highway and carried overturned vehicles into the sea. Heavy flooding also was reported in the Mediterranean resort city of Antalya and in the Aegean port city of Izmir.**

Weather Log - December 11-20th, 2001
According to media reports, Catalonia - in northeast Spain - has remained isolated from the rest of Spain for two days after heavy snow trapped motorists and residents. Further south, in the regions of La Rioja, Castilla and Leon (NW Spain), temperatures fell to minus 10 Celsius (14 degrees Fahrenheit). Spain's Meteorological Institute said the severe cold wave was expected to ease around the 17th and 18th. Heavy snowfalls have caused traffic chaos in Italy, southern France and Spain. Spain suffered the worst problems caused by snow for 15 years, and temperatures in Venice remained well below freezing. Nearly a dozen villages in the remote mountains of northern Corsica remained inaccessible. **In Eastern Europe, snow storms, gale- force winds and heavy rain swept across Greece and Turkey on the 17th, as large parts of Europe remained in the icy grip of winter. Two days of heavy snowfall and temperatures as low as -10C forced the closure of all northern Greek airports, including Thessaloniki international airport, and all schools in northwestern Greece.** In Poland, cold weather has killed 117 people since the beginning of October, officials say.

According to preliminary reports, the world's record high sea level pressure of 1083.8 mb (32.01 inches) recorded at Agata in Siberia on 12/31/1968 appears to have been shattered on December 19, 2001 in Tosontsengel in northwest Mongolia. The city is about 420 miles (680 km) west of the capital city of Ulanbataar. At 2 am local time on the 19th, (18/1800 UTC) the sea level pressure rose to 1085.6 mb (32.06 inches). The town is situated in a protected valley which allows cold air drainage and radiational cooling which leads to high pressure values. At the time of the pressure reading the the temperature was minus 40.5 degrees C (minus 41 degrees F) and there was an almost 4 cm (2 inch) uneven layer of loose dry snow covering the ground completely. The day's low of minus 42.3 degrees C (minus 44 degrees F) and high of minus 32.8 degrees C (minus 27 degrees F) was well below the monthly normal of minus 18 degrees C (0 degrees F).

Now determine summer and winter P and E at Crete (C) and Haifa (H), at sea level:

$$SumPC0 := \sum_{i=0}^{182} PC_i \qquad WinPC0 := \sum_{i=183}^{364} PC_i \qquad SumPCone := \sum_{i=365}^{365+183-1} PC_i$$

$$SumPCthree := \sum_{i=2\cdot 365}^{2\cdot 365+183-1} PC_i \qquad WinPCthree := \sum_{i=2\cdot 365+183}^{(3\cdot 365)-1} PC_i \qquad SumPCfive := \sum_{i=3\cdot 365}^{3\cdot 365+183-1} PC_i$$

$$SumEC0 := \sum_{i=0}^{182} \left(EmmdayC^{\langle 0\rangle}\right)_i \qquad WinEC0 := \sum_{i=183}^{364} \left(EmmdayC^{\langle 0\rangle}\right)_i \qquad SumECone := \sum_{i=365}^{365+183-1} \left(EmmdayC^{\langle 0\rangle}\right)_i$$

$$SumECthree := \sum_{i=2\cdot 365}^{2\cdot 365+183-1} \left(EmmdayC^{\langle 0\rangle}\right)_i \qquad WinECthree := \sum_{i=2\cdot 365+183}^{(3\cdot 365)-1} \left(EmmdayC^{\langle 0\rangle}\right)_i \qquad SumECfive := \sum_{i=3\cdot 365}^{3\cdot 365+183-1} \left(EmmdayC^{\langle 0\rangle}\right)_i$$

$$WinPCone := \sum_{i=365+183}^{(2\cdot 365)-1} PC_i \qquad WinECone := \sum_{i=365+183}^{(2\cdot 365)-1} \left(EmmdayC^{\langle 0\rangle}\right)_i \qquad WinPCfive := \sum_{i=3\cdot 365+183}^{(4\cdot 365)-1} PC_i \qquad WinECfive := \sum_{i=3\cdot 365+183}^{(4\cdot 365)-1} \left(EmmdayC^{\langle 0\rangle}\right)_i$$

$$SumPH0 := \sum_{i=0}^{182} PH_i \qquad WinPH0 := \sum_{i=183}^{364} PH_i \qquad SumPHone := \sum_{i=365}^{365+183-1} PH_i$$

$$SumPHthree := \sum_{i=2\cdot 365}^{2\cdot 365+183-1} PH_i \qquad WinPHthree := \sum_{i=2\cdot 365+183}^{(3\cdot 365)-1} PH_i \qquad SumPHfive := \sum_{i=3\cdot 365}^{3\cdot 365+183-1} PH_i$$

$$SumEH0 := \sum_{i=0}^{182} \left(EmmdayH^{\langle 0\rangle}\right)_i \qquad WinEH0 := \sum_{i=183}^{364} \left(EmmdayH^{\langle 0\rangle}\right)_i \qquad SumEHone := \sum_{i=365}^{365+183-1} \left(EmmdayH^{\langle 0\rangle}\right)_i$$

$$SumEHthree := \sum_{i=2\cdot 365}^{2\cdot 365+183-1} \left(EmmdayH^{\langle 0\rangle}\right)_i \qquad WinEHthree := \sum_{i=2\cdot 365+183}^{(3\cdot 365)-1} \left(EmmdayH^{\langle 0\rangle}\right)_i \qquad SumEHfive := \sum_{i=3\cdot 365}^{3\cdot 365+183-1} \left(EmmdayH^{\langle 0\rangle}\right)_i$$

$$\mathrm{WinPHone} := \sum_{i=365+183}^{(2\cdot 365)-1} PH_i \qquad \mathrm{WinEHone} := \sum_{i=365+183}^{(2\cdot 365)-1} \left(EmmdayH^{\langle 0\rangle}\right)_i \qquad \mathrm{WinPHfive} := \sum_{i=3\cdot 365+183}^{(4\cdot 365)-1} PH_i \qquad \mathrm{WinEHfive} := \sum_{i=3\cdot 365+183}^{(4\cdot 365)-1} \left(EmmdayH^{\langle 0\rangle}\right)_i$$

Calculate stable O-isotope changes

Use Rohling (1999 Paleoceanography) calculation for δe, and hence fractionation between δM and δe:

Resistance coefficient Ratio = 1.0142 after Gonfiantini (1986), but some say this is too large (see Rohling 1999), so we use this to tune the 1-event "control" year rainfall deltas to about -6. I get close to this when I set Ratio = 1.0105 (for cool=6), or Ratio=1.0098 (for cool=10): we settle for 1.010

$$\alpha evap_i := e^{1.137\cdot\left(Ts0_i+273.5\right)^{-2}\,10^{3}-0.4156\cdot\left(Ts0_i+273.15\right)^{-1}-2.0667\cdot 10^{-3}}$$

$$Ratio := 1.010$$

Aegean δe:

$$\delta eAeg_i := \left[\frac{\dfrac{1+\delta sw0\cdot 10^{-3}}{\alpha evap_i} - \dfrac{qa_i}{qs_i}\cdot r0_i\cdot\left(1+\delta a0_i\cdot 10^{-3}\right)}{\left(1-\dfrac{qa_i}{qs_i}\cdot r0_i\right)\cdot Ratio} - 1\right]\cdot 10^{3}$$

Condensation fractionations for the O isotopes:

$$fraccC_i := 1000\left[e^{1.137\cdot\left(TaDC0_i-\Delta TvC_i+273.5\right)^{-2}\,10^{3}-0.4156\cdot\left(TaDC0_i-\Delta TvC_i+273.15\right)^{-1}-2.0667\cdot 10^{-3}} - 1\right]$$

$$fraccH_i := 1000\left[e^{1.137\cdot\left(TaDH0_i-\Delta TvH_i+273.5\right)^{-2}\,10^{3}-0.4156\cdot\left(TaDH0_i-\Delta TvH_i+273.15\right)^{-1}-2.0667\cdot 10^{-3}} - 1\right]$$

Isotope mass balancing between import, evap, and final

$$\delta MendC_i := \frac{\delta a0_i\cdot M0_i + \delta eAeg_i\cdot dayEadwC_i\cdot TtC_i}{MendC_i}$$

Initial rainfall δ values:

$$\delta PC1_i := \delta MendC_i + fraccC_i$$

δPC at **end**point of Rayleigh distillation

$$\delta PCx_i := \delta MendC_i + fraccC_i\cdot\ln\left(fC_i\right) + fraccC_i$$

Calculate change (anomaly) in oxygen isotope values of Aegean surface water (assuming a reset to original value at year end)

$$d\delta sw_i := \frac{\delta sw0 \cdot mld + \delta eAeg_i \cdot \frac{EC_i}{365}}{mld + \frac{EC_i}{365}} - \delta sw0$$

$$\Delta\delta sw0 := \sum_{i=0}^{364} d\delta sw_i \qquad \Delta\delta sw1 := \sum_{i=365}^{(2\cdot 365)-1} d\delta sw_i \qquad \Delta\delta sw3 := \sum_{i=2\cdot 365}^{(3\cdot 365)-1} d\delta sw_i \qquad \Delta\delta sw5 := \sum_{i=3\cdot 365}^{(4\cdot 365)-1} d\delta sw_i$$

$\Delta\delta sw0 = 0.3829$

$\Delta\delta sw1 = 0.4075$ $\quad$ $\Delta\delta sw3 = 0.4652$ $\quad$ $\Delta\delta sw5 = 0.5166$

$\boxed{16a}$

Thus, the final δP is represented by the Natural Lograrithm-weighted mean δPC1 & PCx

And ensure δP only has a value if there is P! Otherwise seasonal means get skewed.

$$\delta PC_i := \begin{cases} \ln\left(\dfrac{e^{\delta PC1_i} + e^{\delta PCx_i}}{2}\right) & \text{if } PC_i > 0 \\ \text{NaN} & \text{otherwise} \end{cases}$$

remaining vapour δa:

$$\delta aendC_i := \begin{cases} \dfrac{1\cdot\delta MendC_i - \left(1 - fC_i\right)\cdot\delta PC_i}{fC_i} & \text{if } PC_i > 0 \\ \delta MendC_i & \text{otherwise} \end{cases}$$

Levantine δe:

$$\delta eLev_i := \left[\frac{\dfrac{1 + \delta sw0\cdot 10^{-3}}{\alpha evap_i} - \dfrac{qaC_i}{qs_i}\cdot rCout_i\cdot\left(1 + \delta aendC_i\cdot 10^{-3}\right)}{\left(1 - \dfrac{qaC_i}{qs_i}\cdot rCout_i\right)\cdot Ratio} - 1\right]\cdot 10^3$$

$$\delta MendH_i := \frac{\left(\delta aendC_i\cdot MendC_i\cdot fC_i\cdot\dfrac{\mu\ VHm_i}{LeH - LeC} + \delta eLev_i\cdot dayEadwH_i\right)\cdot\left(TtH_i - TtC_i\right)}{MendH_i\ spacH}$$

Initial rainfall δ values:

$$\delta PH1_i := \delta MendH_i + fraccH_i$$

δPH at **end**point of Rayleigh distillation

$$\delta PHx_i := \delta MendH_i + fraccH_i\cdot\ln\left(fH_i\right) + fraccH_i$$

Now the entire δP is represented by the Natural lograrithm-weighted mean δPC

And ensure δP only has a value if there is P! Otherwise seasonal means get skewed.

$$\delta PH_i := \begin{cases} \ln\left(\dfrac{e^{\delta PH1_i} + e^{\delta PHx_i}}{2}\right) & \text{if } PH_i > 0 \\ \text{NaN} & \text{otherwise} \end{cases}$$

mean rainfall δP values for Crete:

Disable the functions when the code flags that "Sub" in the loops is empty

$$\text{ann}\delta\text{PC0} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PC}, 0, 365-1, 0, 0) \\ y \leftarrow \text{mean}(\text{filterNaN}(\text{Sub})) \end{array} \right.$$

$$\text{sum}\delta\text{PC0} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PC}, 0, 182, 0, 0) \\ y \leftarrow \text{mean}(\text{filterNaN}(\text{Sub})) \end{array} \right.$$

$$\text{win}\delta\text{PC0} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PC}, 183, 365-1, 0, 0) \\ y \leftarrow \text{mean}(\text{Sub}) \end{array} \right.$$

$$\text{ann}\delta\text{PCone} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PC}, 365, 2\cdot 365-1, 0, 0) \\ y \leftarrow \text{mean}(\text{filterNaN}(\text{Sub})) \end{array} \right.$$

$$\text{win}\delta\text{PCone} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PC}, 365+183, 2\cdot 365-1, 0, 0) \\ y \leftarrow \text{mean}(\text{Sub}) \end{array} \right.$$

$$\text{sum}\delta\text{PCone} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PC}, 365, 365+183-1, 0, 0) \\ y \leftarrow \text{mean}(\text{filterNaN}(\text{Sub})) \end{array} \right.$$

$$\text{win}\delta\text{PCthree} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PC}, 2\cdot 365+183, 3\cdot 365-1, 0, 0) \\ y \leftarrow \text{mean}(\text{Sub}) \end{array} \right.$$

$$\text{ann}\delta\text{PCthree} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PC}, 2\cdot 365, 3\cdot 365-1, 0, 0) \\ y \leftarrow \text{mean}(\text{filterNaN}(\text{Sub})) \end{array} \right.$$

$$\text{sum}\delta\text{PCthree} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PC}, 2\cdot 365, 2\cdot 365+183-1, 0, 0) \\ y \leftarrow \text{mean}(\text{filterNaN}(\text{Sub})) \end{array} \right.$$

$$\text{ann}\delta\text{PCfive} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PC}, 3\cdot 365, 4\cdot 365-1, 0, 0) \\ y \leftarrow \text{mean}(\text{filterNaN}(\text{Sub})) \end{array} \right.$$

$$\text{win}\delta\text{PCfive} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PC}, 3\cdot 365+183, 4\cdot 365-1, 0, 0) \\ y \leftarrow \text{mean}(\text{Sub}) \end{array} \right.$$

$$\text{sum}\delta\text{PCfive} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PC}, 3\cdot 365, 3\cdot 365+183-1, 0, 0) \\ y \leftarrow \text{mean}(\text{filterNaN}(\text{Sub})) \end{array} \right.$$

$\text{ann}\delta\text{PC0} = -6.04$

$\text{ann}\delta\text{PCone} = -6.172$

$\text{ann}\delta\text{PCthree} = -6.398$

$\text{ann}\delta\text{PCfive} = -6.639$

Apparent interannual amount effect:

$$\Delta C3 := \frac{\text{ann}\delta\text{PCthree} - \text{ann}\delta\text{PCone}}{\text{WinPCthree} + \text{SumPCthree} - \text{WinPCone} - \text{SumPCone}} \cdot 100$$

$$\Delta C5 := \frac{\text{ann}\delta\text{PCfive} - \text{ann}\delta\text{PCone}}{\text{WinPCfive} + \text{SumPCfive} - \text{WinPCone} - \text{SumPCone}} \cdot 100$$

$$\text{me}\Delta C := \frac{\Delta C3 + \Delta C5}{2}$$

$\Delta C3 = -0.122$ $\Delta C5 = -0.159$ permil per 100 mm $\text{me}\Delta C = -0.141$

$\Delta\text{SSTnone} := \text{annchange0event} - \text{annchange}$
$\Delta\text{SSTthree} := \text{annchange3event} - \text{annchange}$
$\Delta\text{SSTfive} := \text{annchange5event} - \text{annchange}$

$\Delta\text{SSTnone} = 0.719$
$\Delta\text{SSTthree} = -1.753$
$\Delta\text{SSTfive} = -3.273$

$\text{cool} = 9$ $\text{maxheiC} = 5000$ $\text{maxheiH} = 6500$

Mean rainfall δP values for Levant (Haifa): *Disable the functions when the code flags that "Sub" in the loops is empty*

$$\text{ann}\delta\text{PH0} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PH}, 0, 365-1, 0, 0) \\ \text{y} \leftarrow \text{mean}(\text{filterNaN}(\text{Sub})) \end{array} \right.$$

$$\text{sum}\delta\text{PH0} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PH}, 0, 182, 0, 0) \\ \text{y} \leftarrow \text{mean}(\text{filterNaN}(\text{Sub})) \end{array} \right.$$

$$\text{win}\delta\text{PH0} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PH}, 183, 365-1, 0, 0) \\ \text{y} \leftarrow \text{mean}(\text{Sub}) \end{array} \right.$$

$$\text{ann}\delta\text{PHone} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PH}, 365, 2\cdot 365-1, 0, 0) \\ \text{y} \leftarrow \text{mean}(\text{filterNaN}(\text{Sub})) \end{array} \right.$$

$$\text{win}\delta\text{PHone} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PH}, 365+183, 2\cdot 365-1, 0, 0) \\ \text{y} \leftarrow \text{mean}(\text{Sub}) \end{array} \right.$$

$$\text{sum}\delta\text{PHone} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PH}, 365, 365+183-1, 0, 0) \\ \text{y} \leftarrow \text{mean}(\text{filterNaN}(\text{Sub})) \end{array} \right.$$

$$\text{ann}\delta\text{PHthree} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PH}, 2\cdot 365, 3\cdot 365-1, 0, 0) \\ \text{y} \leftarrow \text{mean}(\text{filterNaN}(\text{Sub})) \end{array} \right.$$

$$\text{win}\delta\text{PHthree} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PH}, 2\cdot 365+183, 3\cdot 365-1, 0, 0) \\ \text{y} \leftarrow \text{mean}(\text{Sub}) \end{array} \right.$$

$$\text{sum}\delta\text{PHthree} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PH}, 2\cdot 365, 2\cdot 365+183-1, 0, 0) \\ \text{y} \leftarrow \text{mean}(\text{filterNaN}(\text{Sub})) \end{array} \right.$$

$$\text{ann}\delta\text{PHfive} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PH}, 3\cdot 365, 4\cdot 365-1, 0, 0) \\ \text{y} \leftarrow \text{mean}(\text{filterNaN}(\text{Sub})) \end{array} \right.$$

$$\text{win}\delta\text{PHfive} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PH}, 3\cdot 365+183, 4\cdot 365-1, 0, 0) \\ \text{mean}(\text{Sub}) \end{array} \right.$$

$$\text{sum}\delta\text{PHfive} := \left| \begin{array}{l} \text{Sub} \leftarrow \text{submatrix}(\delta\text{PH}, 3\cdot 365, 3\cdot 365+183-1, 0, 0) \\ \text{y} \leftarrow \text{mean}(\text{filterNaN}(\text{Sub})) \end{array} \right.$$

$\text{ann}\delta\text{PH0} = -5.351$

$\text{ann}\delta\text{PHone} = -5.658$

$\text{ann}\delta\text{PHthree} = -6.453$

$\text{ann}\delta\text{PHfive} = -7.097$

Apparent interannual amount effect:

$$\Delta\text{H3} := \frac{\text{ann}\delta\text{PHthree} - \text{ann}\delta\text{PHone}}{\text{WinPHthree} + \text{SumPHthree} - \text{WinPHone} - \text{SumPHone}}\cdot 100$$

$$\Delta\text{H5} := \frac{\text{ann}\delta\text{PHfive} - \text{ann}\delta\text{PHone}}{\text{WinPHfive} + \text{SumPHfive} - \text{WinPHone} - \text{SumPHone}}\cdot 100$$

$$\text{me}\Delta\text{H} := \frac{\Delta\text{H3} + \Delta\text{H5}}{2}$$

$\Delta\text{H3} = -0.471$ $\Delta\text{H5} = -0.529$ permil per 100 mm $\text{me}\Delta\text{H} = -0.5$

Holocene runs, using GISP2 K± series to identify periods with enhanced-frequency northerly air outbreaks (Rohling et al., 2002 Clim. Dyn.)

These runs use the cut-offs in thef GISP2 K+ ion series to identify periods with certain frequencies of northerly outbreaks. They allow for periods with fewer events than today by setting the control scenario to the 1-event years, and the 0-event years then are "more optimal" than today.

$yrs := \mathrm{length}(GISPt)$ $\quad yrs = 3.777 \times 10^{3}$ $\quad \max(GISPt) = 1.134 \times 10^{4}$

$h := 0 .. yrs - 1$ $\quad x_h := GISPKraw_h$ $\quad tini := -GISPt$

$p95 := \mathrm{percentile}(GISPKraw, 0.95)$ $\quad p95 = 1.661$

$p80 := \mathrm{percentile}(GISPKraw, 0.8)$ $\quad p80 = 1.069$

$p68 := \mathrm{percentile}(GISPKraw, 0.68)$ $\quad p68 = 0.89$

$p50 := \mathrm{percentile}(GISPKraw, 0.50)$ $\quad p50 = 0.738$

Categorising event frequencies from K+ scores:

$$yini_h := \left| \begin{array}{ll} p_h \leftarrow 1 \ \text{if} \ x_h > 0.74 & \text{50th percentile} \\ p_h \leftarrow 3 \ \text{if} \ x_h > 1.07 & \text{80th percentile} \\ p_h \leftarrow 5 \ \text{if} \ x_h > 1.66 & \text{95th percentile} \\ p_h \leftarrow 0 \ \text{otherwise} & \\ p_h & \end{array} \right.$$

$MAT := \mathrm{augment}(tini, yini)$

$t := \left(\mathrm{csort}(MAT, 0)^{\langle 0 \rangle}\right)$

$y0 := \left(\mathrm{csort}(MAT, 0)^{\langle 1 \rangle}\right)$

$y := y0$

Set counter value (h) for present-day value $\quad pres \equiv 3776$

Check that present-day is a 1-event year $\quad y_{pres} = 1$ $\quad$ All normalisations are OK as long as this value is equal to 1

First, a little **calculation of "negative degree days" at a selected altitude "alt" in Crete**. It assumes that negative degree days are indicative of the **potential** for snowfield development.

Set preferred altitude $\quad alt := 2000$

$$TC_{alt_i} := \left| \begin{array}{l} TaC0_i - \frac{alt}{1000} \cdot LR \ \text{if} \ TaC0_i - \frac{alt}{1000} \cdot LR \le 0 \\ 0 \ \text{otherwise} \end{array} \right.$$

$$SC0 := \sum_{i=183}^{(365)-1} TC_{alt_i} \qquad SC1 := \sum_{i=365+183}^{(2\cdot 365)-1} TC_{alt_i} \qquad SC3 := \sum_{i=2\cdot 365+183}^{(3\cdot 365)-1} TC_{alt_i} \qquad SC5 := \sum_{i=3\cdot 365+183}^{(4\cdot 365)-1} TC_{alt_i}$$

$SC0 = 0$ $\quad SC1 = -77$ $\quad SC3 = -111.638$ $\quad SC5 = -231$

Now **make time-series using 0,1,3,5 to select the year characteristics as defined above**.

$$\Delta SST_h := \begin{cases} \Delta SSTnone & \text{if } -0.01 < y_h < 0.01 \\ 0 & \text{if } 0.99 < y_h < 1.01 \\ \Delta SSTthree & \text{if } 2.99 < y_h < 3.01 \\ \Delta SSTfive & \text{if } 4.99 < y_h < 5.01 \end{cases}$$

$$winPC_h := \begin{cases} WinPC0 & \text{if } -0.01 < y_h < 0.01 \\ WinPCone & \text{if } 0.99 < y_h < 1.01 \\ WinPCthree & \text{if } 2.99 < y_h < 3.01 \\ WinPCfive & \text{if } 4.99 < y_h < 5.01 \end{cases}$$

$$\Delta SSS_h := \begin{cases} \Delta SSS0 - \Delta SSS1 & \text{if } -0.01 < y_h < 0.01 \\ 0 & \text{if } 0.99 < y_h < 1.01 \\ \Delta SSS3 - \Delta SSS1 & \text{if } 2.99 < y_h < 3.01 \\ \Delta SSS5 - \Delta SSS1 & \text{if } 4.99 < y_h < 5.01 \end{cases}$$

$$winPH_h := \begin{cases} WinPH0 & \text{if } -0.01 < y_h < 0.01 \\ WinPHone & \text{if } 0.99 < y_h < 1.01 \\ WinPHthree & \text{if } 2.99 < y_h < 3.01 \\ WinPHfive & \text{if } 4.99 < y_h < 5.01 \end{cases}$$

$$win\delta PC_h := \begin{cases} win\delta PC0 & \text{if } -0.01 < y_h < 0.01 \\ win\delta PCone & \text{if } 0.99 < y_h < 1.01 \\ win\delta PCthree & \text{if } 2.99 < y_h < 3.01 \\ win\delta PCfive & \text{if } 4.99 < y_h < 5.01 \end{cases}$$

$$win\delta PH_h := \begin{cases} win\delta PH0 & \text{if } -0.01 < y_h < 0.01 \\ win\delta PHone & \text{if } 0.99 < y_h < 1.01 \\ win\delta PHthree & \text{if } 2.99 < y_h < 3.01 \\ win\delta PHfive & \text{if } 4.99 < y_h < 5.01 \end{cases}$$

$$ann\delta PH_h := \begin{cases} ann\delta PH0 & \text{if } -0.01 < y_h < 0.01 \\ ann\delta PHone & \text{if } 0.99 < y_h < 1.01 \\ ann\delta PHthree & \text{if } 2.99 < y_h < 3.01 \\ ann\delta PHfive & \text{if } 4.99 < y_h < 5.01 \end{cases}$$

$$sumEC_h := \begin{cases} SumEC0 & \text{if } -0.01 < y_h < 0.01 \\ SumECone & \text{if } 0.99 < y_h < 1.01 \\ SumECthree & \text{if } 2.99 < y_h < 3.01 \\ SumECfive & \text{if } 4.99 < y_h < 5.01 \end{cases}$$

$$winEC_h := \begin{cases} WinEC0 & \text{if } -0.01 < y_h < 0.01 \\ WinECone & \text{if } 0.99 < y_h < 1.01 \\ WinECthree & \text{if } 2.99 < y_h < 3.01 \\ WinECfive & \text{if } 4.99 < y_h < 5.01 \end{cases}$$

$$sumEH_h := \begin{cases} SumEH0 & \text{if } -0.01 < y_h < 0.01 \\ SumEHone & \text{if } 0.99 < y_h < 1.01 \\ SumEHthree & \text{if } 2.99 < y_h < 3.01 \\ SumEHfive & \text{if } 4.99 < y_h < 5.01 \end{cases}$$

$$winEH_h := \begin{cases} WinEH0 & \text{if } -0.01 < y_h < 0.01 \\ WinEHone & \text{if } 0.99 < y_h < 1.01 \\ WinEHthree & \text{if } 2.99 < y_h < 3.01 \\ WinEHfive & \text{if } 4.99 < y_h < 5.01 \end{cases}$$

$$NDDC_h := \begin{cases} SC0 & \text{if } -0.01 < y_h < 0.01 \\ SC1 & \text{if } 0.99 < y_h < 1.01 \\ SC3 & \text{if } 2.99 < y_h < 3.01 \\ SC5 & \text{if } 4.99 < y_h < 5.01 \end{cases}$$

$$annTaH0 := \frac{1}{365}\cdot\sum_{i=0}^{364}\left(TaH0^{\langle 0\rangle}\right)_i \qquad winTaH0 := \frac{1}{183}\cdot\sum_{i=183}^{364}\left(TaH0^{\langle 0\rangle}\right)_i$$

$$annTaHone := \frac{1}{365}\cdot\sum_{i=365}^{(2\cdot 365)-1}\left(TaH0^{\langle 0\rangle}\right)_i \qquad winTaHone := \frac{1}{183}\cdot\sum_{i=365+183}^{(2\cdot 365)-1}\left(TaH0^{\langle 0\rangle}\right)_i$$

$$annTaHthree := \frac{1}{365}\cdot\sum_{i=2\cdot 365}^{(3\cdot 365)-1}\left(TaH0^{\langle 0\rangle}\right)_i \qquad winTaHthree := \frac{1}{183}\cdot\sum_{i=2\cdot 365+183}^{(3\cdot 365)-1}\left(TaH0^{\langle 0\rangle}\right)_i \qquad annTaHfive := \frac{1}{365}\cdot\sum_{i=3\cdot 365}^{(4\cdot 365)-1}\left(TaH0^{\langle 0\rangle}\right)_i$$

$$winTaHfive := \frac{1}{183}\cdot\sum_{i=3\cdot 365+183}^{(4\cdot 365)-1}\left(TaH0^{\langle 0\rangle}\right)_i$$

$$annTaH_h := \left|\begin{array}{l} annTaH0 \text{ if } -0.01 < y_h < 0.01 \\ annTaHone \text{ if } 0.99 < y_h < 1.01 \\ annTaHthree \text{ if } 2.99 < y_h < 3.01 \\ annTaHfive \text{ if } 4.99 < y_h < 5.01 \end{array}\right. \qquad winTaH_h := \left|\begin{array}{l} winTaH0 \text{ if } -0.01 < y_h < 0.01 \\ winTaHone \text{ if } 0.99 < y_h < 1.01 \\ winTaHthree \text{ if } 2.99 < y_h < 3.01 \\ winTaHfive \text{ if } 4.99 < y_h < 5.01 \end{array}\right.$$

$$sumPC_h := \left|\begin{array}{l} SumPC0 \text{ if } -0.01 < y_h < 0.01 \\ SumPCone \text{ if } 0.99 < y_h < 1.01 \\ SumPCthree \text{ if } 2.99 < y_h < 3.01 \\ SumPCfive \text{ if } 4.99 < y_h < 5.01 \end{array}\right. \qquad sumPH_h := \left|\begin{array}{l} SumPH0 \text{ if } -0.01 < y_h < 0.01 \\ SumPHone \text{ if } 0.99 < y_h < 1.01 \\ SumPHthree \text{ if } 2.99 < y_h < 3.01 \\ SumPHfive \text{ if } 4.99 < y_h < 5.01 \end{array}\right.$$

$$ann\delta PC_h := \left|\begin{array}{l} ann\delta PC0 \text{ if } -0.01 < y_h < 0.01 \\ ann\delta PCone \text{ if } 0.99 < y_h < 1.01 \\ ann\delta PCthree \text{ if } 2.99 < y_h < 3.01 \\ ann\delta PCfive \text{ if } 4.99 < y_h < 5.01 \end{array}\right.$$

$$\Delta\delta sw_h := \left|\begin{array}{l} \Delta\delta sw0 - \Delta\delta sw1 \text{ if } -0.01 < y_h < 0.01 \\ 0 \text{ if } 0.99 < y_h < 1.01 \\ \Delta\delta sw3 - \Delta\delta sw1 \text{ if } 2.99 < y_h < 3.01 \\ \Delta\delta sw5 - \Delta\delta sw1 \text{ if } 4.99 < y_h < 5.01 \end{array}\right.$$

Define annual values

$$annPC_h := sumPC_h + winPC_h \qquad annPH_h := sumPH_h + winPH_h \qquad annEC_h := sumEC_h + winEC_h \qquad annEH_h := sumEH_h + winEH_h$$

$$annPCone := SumPCone + WinPCone \qquad annPHone := SumPHone + WinPHone$$

Determine Haifa carbonate O-isotopes (using δP and Temp - both for winter because that's when the precip happens. Hence, this calculation does NOT account for effects of groundwater evaporation)

$$win\delta PHcarb := \left(win\delta PH - win\delta PH_{pres}\right) - \left(winTaH - winTaH_{pres}\right)\cdot 0.25$$

Determine smoothed records to approximate recording in natural archives

Set smoothing factors (y): $sm1 \equiv 50$ $sm2 := 200$ $annTaHsm := ksmooth(t, annTaH, sm1)$

$\Delta SSTsm := ksmooth(t, \Delta SST, sm2)$
$\Delta SSSsm := ksmooth(t, \Delta SSS, sm2)$
$\Delta\delta swsm := ksmooth(t, \Delta\delta sw, sm2)$

$winPCsm := ksmooth(t, winPC, sm1)$

$ann\delta PCsm := ksmooth(t, win\delta PC, sm1)$
$win\delta PHsm := ksmooth(t, win\delta PH, sm1)$

$winPHsm := ksmooth(t, winPH, sm1)$

$sumECsm := ksmooth(t, sumEC, sm1)$
$winECsm := ksmooth(t, winEC, sm1)$

$annPCsm := ksmooth(t, annPC, sm1)$
$annECsm := ksmooth(t, annEC, sm1)$

$winEHsm := ksmooth(t, winEH, sm1)$

$ann\delta PHsm := ksmooth(t, ann\delta PH, sm1)$
$sumEHsm := ksmooth(t, sumEH, sm1)$

$annPHsm := ksmooth(t, annPH, sm1)$
$annEHsm := ksmooth(t, annEH, sm1)$

Although these three anomalies are cumulative over the 3 years of each dt step, we are interested in their annual anomalies, which is what we'd record if we assume the system rests them each year.

$FCsm := ksmooth(t, sumEC + sumPC + winEC + winPC, sm1)$
$FHsm := ksmooth(t, sumEH + sumPH + winEH + winPH, sm1)$

$win\delta PHsm3 := ksmooth(t, ann\delta PH, sm1)$
$win\delta PHcarbsm3 := ksmooth(t, win\delta PHcarb, 20)$

$winTaHsm := ksmooth(t, win\delta PHcarb, sm1)$

Some records we want in anomalies, relative to present value:

$\Delta SSTsm_{norm} := \Delta SSTsm - \Delta SSTsm_{pres}$

$annTaHsm_{norm} := annTaHsm - annTaHsm_{pres}$

$win\delta PHsm_{norm} := win\delta PHsm3 - win\delta PHsm3_{pres}$

$\Delta SSSsm_{norm} := \Delta SSSsm - \Delta SSSsm_{pres}$

$\Delta\delta swsm_{norm} := \Delta\delta swsm - \Delta\delta swsm_{pres}$

$win\delta PHcarbsm_{norm} := win\delta PHcarbsm3 - win\delta PHcarbsm3_{pres}$

$winTaHsm_{norm} := winTaHsm - winTaHsm_{pres}$

carbonate O-isotope anomaly for 3k event re. to present:

$\Delta win\delta carbsm_{3k} := win\delta PHcarbsm_{norm_{2775}} - win\delta PHcarbsm_{norm_{pres}}$

$\Delta win\delta carbsm_{3k} = -0.511$

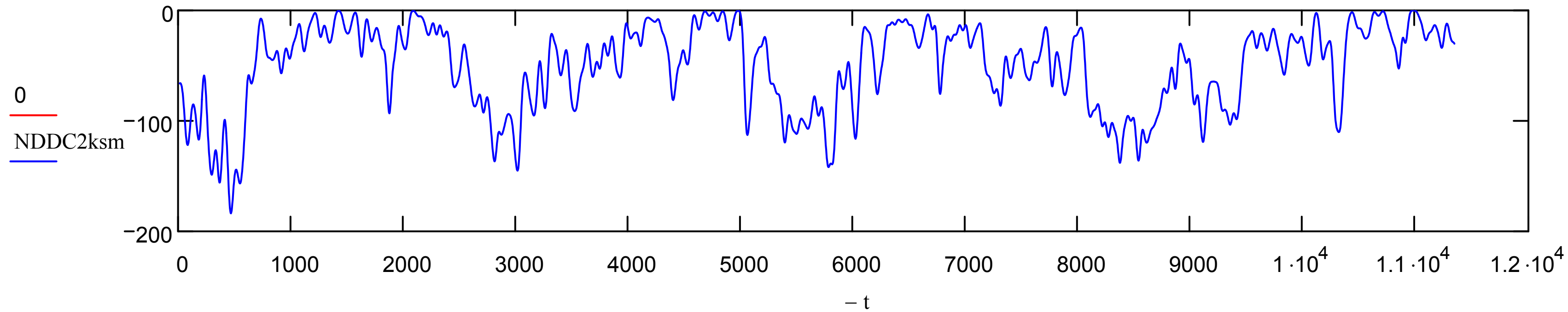
Plots of results

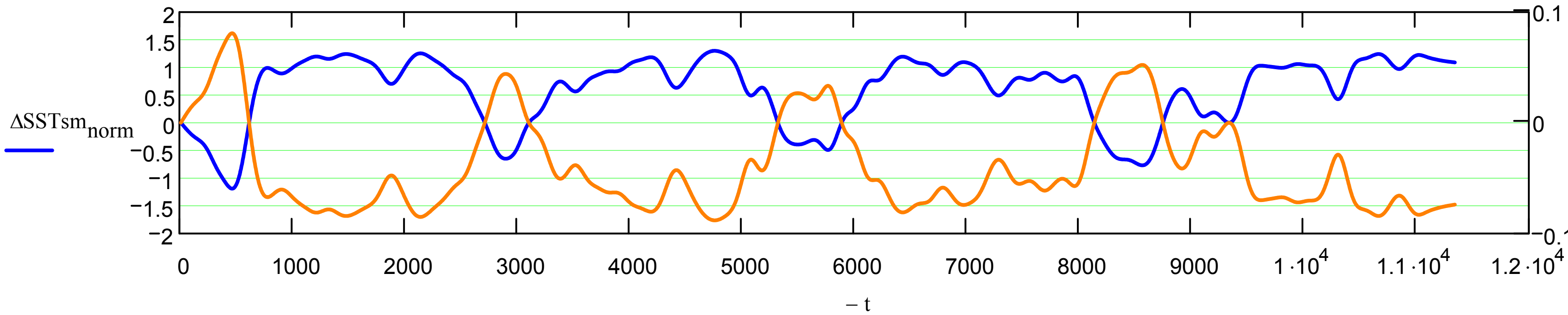
Number of degree days record is multi-decadally smoothed to allow for longevity of snow fields (potential to exist through summers)

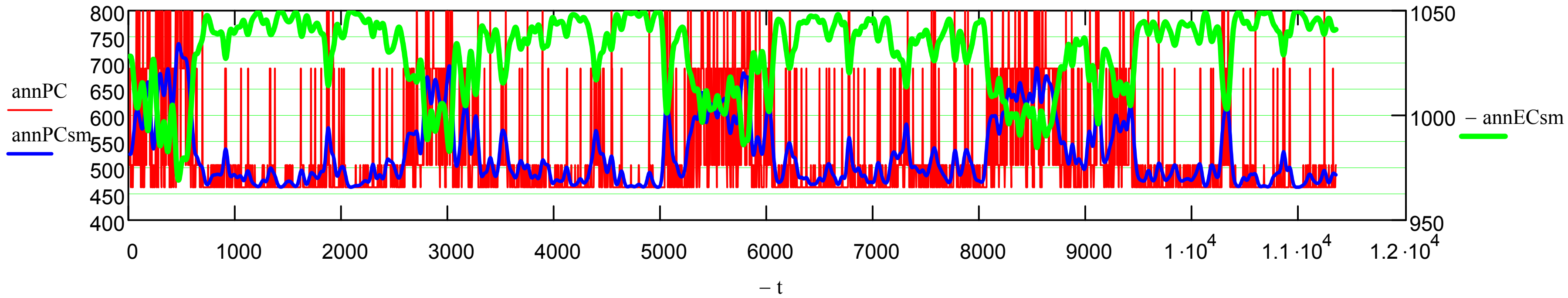
$NDDC2ksm := ksmooth(t, NDDC, sm1)$

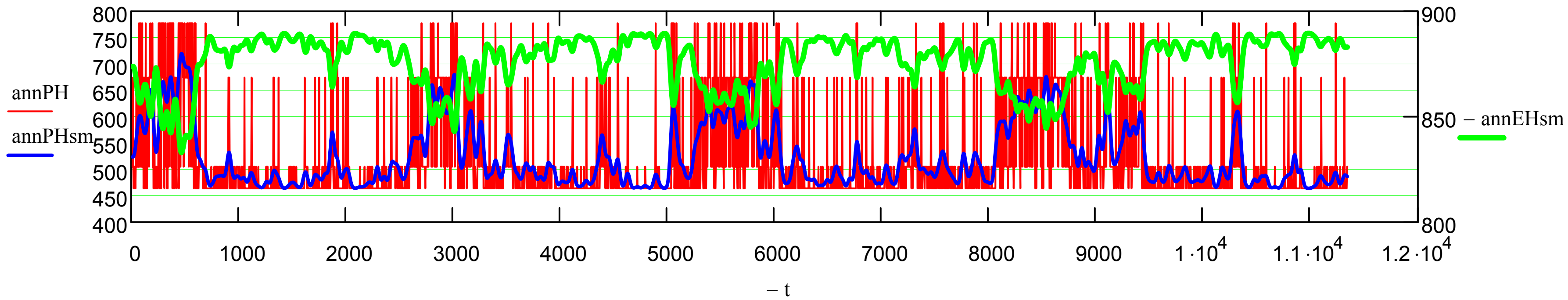
$alt = 2000$

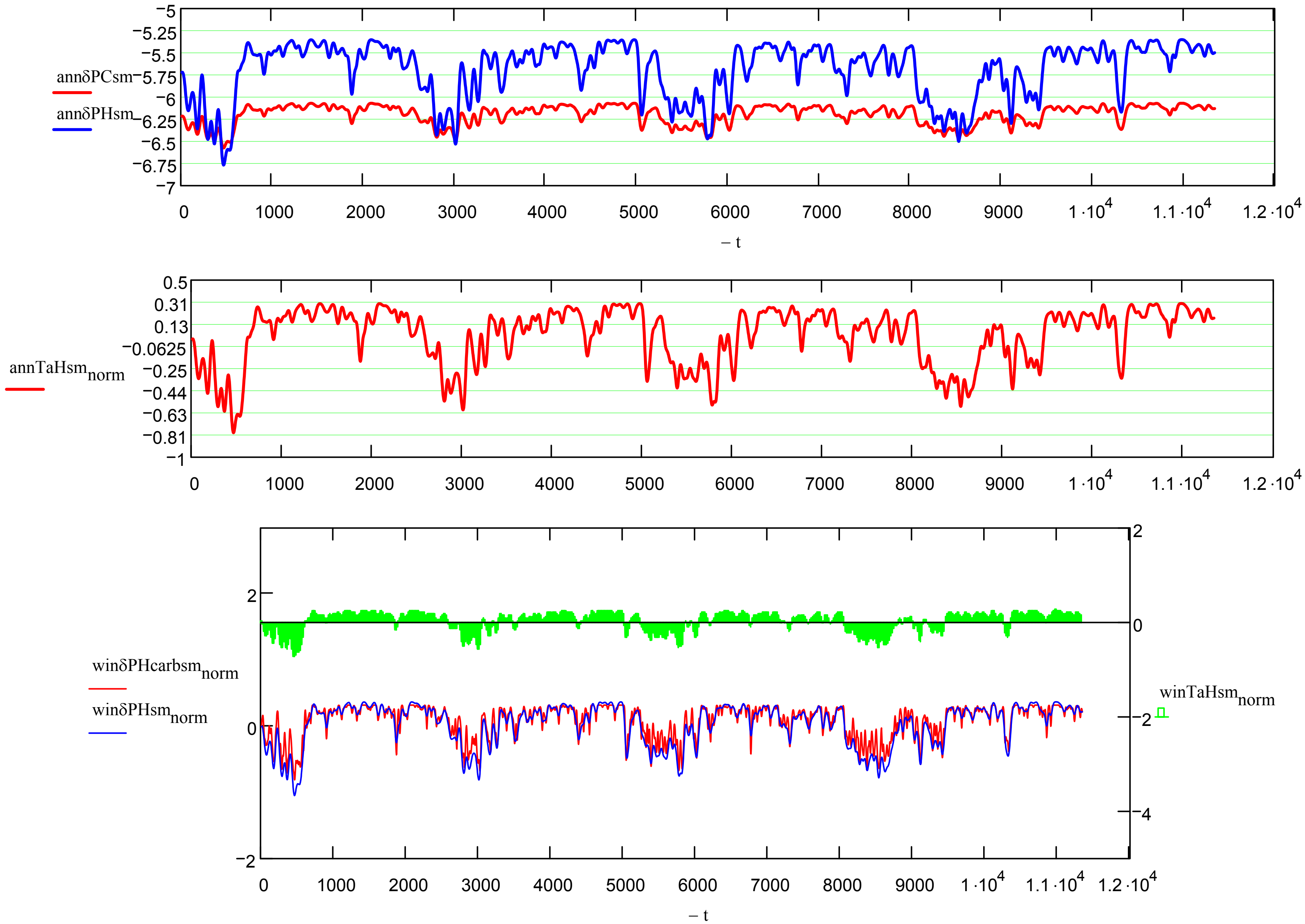
$NDDC2ksm_{pres} = -66.192$




ΔSSTsm_norm
ΔSSSsm_norm
−t


annPC
annPCsm
− annECsm
−t

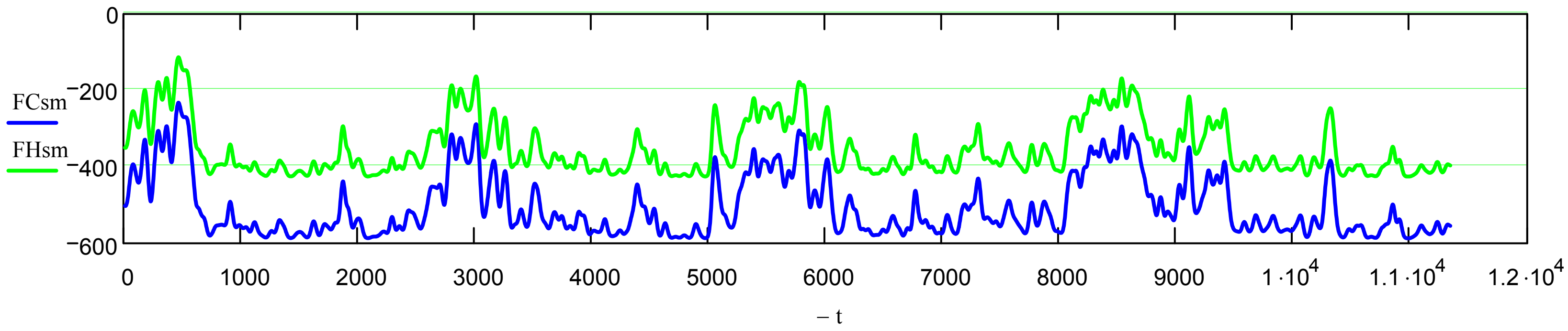

annPH
annPHsm
− annEHsm
−t



annδPCsm
annδPHsm
−5
−5.25
−5.5
−5.75
−6
−6.25
−6.5
−6.75
−7
0
1000
2000
3000
4000
5000
6000
7000
8000
9000
$1 \cdot 10^4$
$1.1 \cdot 10^4$
$1.2 \cdot 10^4$
− t
$annTaHsm_{norm}$
0.5
0.31
0.13
−0.0625
−0.25
−0.44
−0.63
−0.81
−1
$win\delta PHcarbsm_{norm}$
$win\delta PHsm_{norm}$
$winTaHsm_{norm}$
2
0
−2
−4

Lake changes

We can see that the annual freshwater budgets per single square meter are always negative in both locations. So to have sustained lakes, both locations need catchment areas that are larger than the areas of the lakes themselves:



Cretan lake that is at current steady state, must have a cacthment-to-lakearea ratio that compensates for the offset between E and P.

lake area: catchment area

$LA = 1$ $CA = \Phi - 1$

We grossly simplify by assuming that all P immediately soaks deep into the soil, and thus that in-catchment wE=0. wE only has value over the lake itself.

$sE = 1 \cdot sumEC$ $sE + wE + \Phi \cdot wP = 0$

$wE = 1 \cdot winEC$

$wP = \Phi \cdot winPC$

steady state:

$$\Phi setC := \begin{cases} \dfrac{sumEC_{pres} + winEC_{pres}}{(-winPC)_{pres} - sumPC_{pres}} & \text{if } \dfrac{sumEC_{pres} + winEC_{pres}}{(-winPC)_{pres} - sumPC_{pres}} \geq 1 \\ NaN & \text{otherwise} \end{cases}$$

$\Phi setC = 2.04242$

Φset is determined in such a way that the lake is at steady state during a "control" year (a 1-event year)

or override: $\Phi setC := 40.5$

Lake-level change are in **m/y**

relationship used: $depth = \tan(\gamma) \cdot \dfrac{dia}{2}$

Lake-wall slope in degrees: $deg \equiv 1$

Initial lake depth: Set once. $depth0 := 30$

$\gamma := deg \cdot \dfrac{\pi}{180}$ in radians $\tan(\gamma) = 0.017$

Lake diameter (m): $dia := 2 \cdot \dfrac{depth0}{\tan(\gamma)}$ $LA0 := \left(\dfrac{dia}{2}\right)^2 \cdot \pi$

diameter of lake: $dia = 3437$

initial area of lake: $LA0 = 9.28 \times 10^6$

area of lake+catchment: $(\Phi setC) \cdot LA0 = 1.9 \times 10^7$

Levantine lake that is at current steady state, must have a cacthment-to-lakearea ratio that compensates for the offset between E and P.

lake area: catchment area

$LA = 1$ $CA = \Phi - 1$

steady state:

$$\Phi setH := \begin{cases} \dfrac{sumEH_{pres} + winEH_{pres}}{(-winPH)_{pres} - sumPH_{pres}} & \text{if } \dfrac{sumEH_{pres} + winEH_{pres}}{(-winPH)_{pres} - sumPH_{pres}} \geq 1 \\ NaN & \text{otherwise} \end{cases}$$

$\Phi setH = 1.73294$

We grossly simplify by assuming that all P immediately soaks deep into the soil, and thus that in-catchment wE=0. wE only has value over the lake itself.

$sE = 1 \cdot sumEC$ $sE + wE + \Phi \cdot wP = 0$

$wE = 1 \cdot winEC$

$wP = \Phi \cdot winPC$

Φset is determined in such a way that the lake is at steady state during a "control" year (a 1-event year)

or override: $\Phi setH := 40$

Lake-level changes are in **m/y**

Lake-wall slope in degrees: **Initial lake depth:** Set once.

$degH \equiv 1$ $depth0H := 30$ $\gamma H := degH \cdot \dfrac{\pi}{180}$ in radians $\tan(\gamma H) = 0.017$

Lake diameter (m): $diaH := 2 \cdot \dfrac{depth0H}{\tan(\gamma H)}$ $LA0H := \left(\dfrac{diaH}{2}\right)^2 \cdot \pi$

diameter of lake: $diaH = 3437$

initial area of lake: $LA0H = 9.28 \times 10^6$

area of lake+catchment: $(\Phi setH) \cdot LA0H = 1.61 \times 10^7$

For these parameters, calculate lake-level changes through time, in the following:

$$
\begin{aligned}
LLC :=\ & LL_0 \leftarrow 0 \\
& LA_0 \leftarrow LA0 \\
& CA \leftarrow (\Phi setC - 1)\cdot LA0 \\
& \Delta LL_0 \leftarrow \frac{sumEC_0 + winEC_0 + \Phi setC\cdot winPC_0}{1000}\cdot dt \\
& \text{for } p \in 1..pres \\
& \quad LA_p \leftarrow \left(\sqrt{\frac{LA_{p-1}}{\pi}} + \frac{\Delta LL_{p-1}}{\tan(\gamma)}\right)^2\cdot\pi \\
& \quad Ldia_p \leftarrow 2\cdot\sqrt{\frac{LA_p}{\pi}} \\
& \quad \Phi_p \leftarrow \frac{CA + 1}{LA_p} \\
& \quad \Delta LL_p \leftarrow \frac{sumEC_p + winEC_p + \Phi_p\cdot winPC_p}{1000}\cdot dt \\
& \quad LL_p \leftarrow \Delta LL_p + LL_{p-1}
\end{aligned}
$$

Cretan lake

NB. All E and P values are given in mm, and we want answers for LL in m; hence divide by 1000

$LLC_{pres} = -7.74$

$t_{pres} = -15 \quad LLC_{pres} = -7.74$

$\Delta LLC_h := LLC_h - LLC_{pres}$

$\Delta LLCsm := ksmooth(t, \Delta LLC, sm1)$

Repeat values:

$cool = 9$

$Vinc = 10$

$dia = 3437$

$deg = 1$

$depth0 = 30$

Note: to determine dt

$g := 0..pres - 1$

$tdif_g := tini_g - tini_{g+1}$

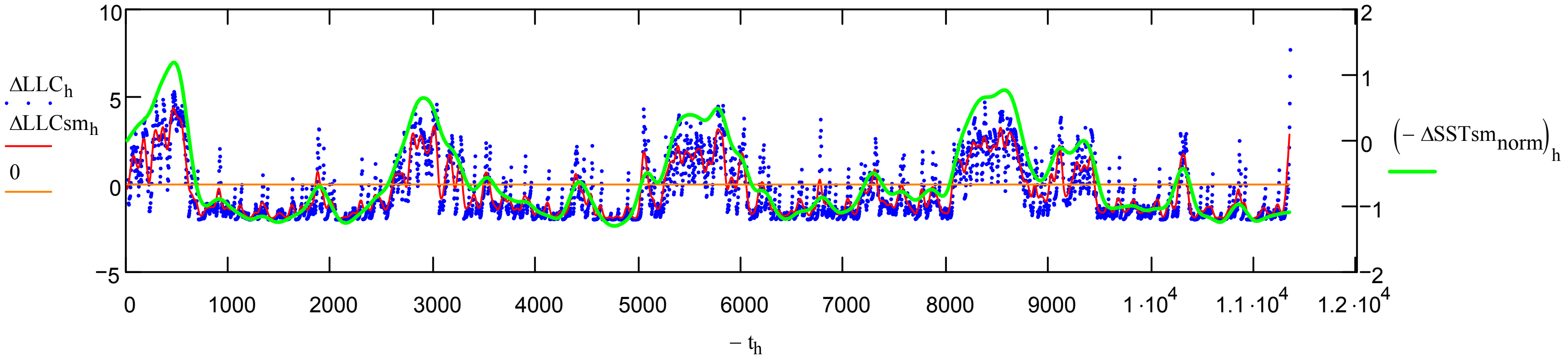
tdif =

	0
0	3
1	3
2	3
3	3
4	3
5	3
6	3
7	3
8	3
9	3

set dt manually based on tdif determined above

$dt \equiv 3$

Lake-level changes relative to Present



Levantine lake

$$
LLH := \left|
\begin{array}{l}
LL_0 \leftarrow 0 \\
LA_0 \leftarrow LA0H \\
CA \leftarrow (\Phi setH - 1) \cdot LA0H \\
\Delta LL_0 \leftarrow \dfrac{sumEH_0 + winEH_0 + \Phi setH \cdot winPH_0}{1000} \cdot dt \\
\text{for } p \in 1 .. pres \\
\quad \left|
\begin{array}{l}
LA_p \leftarrow \left(\sqrt{\dfrac{LA_{p-1}}{\pi}} + \dfrac{\Delta LL_{p-1}}{\tan(\gamma H)} \right)^2 \cdot \pi \\
Ldia_p \leftarrow 2 \cdot \sqrt{\dfrac{LA_p}{\pi}} \\
\Phi_p \leftarrow \dfrac{CA + 1}{LA_p} \\
\Delta LL_p \leftarrow \dfrac{sumEH_p + winEH_p + \Phi_p \cdot winPH_p}{1000} \cdot dt \\
LL_p \leftarrow \Delta LL_p + LL_{p-1}
\end{array}
\right.
\end{array}
\right.
$$

NB. All E and P values are given in mm, and we want answers for LL in m; hence divide by 1000

Repeat values:

$cool = 9$

$Vinc = 10$

$diaH = 3437$

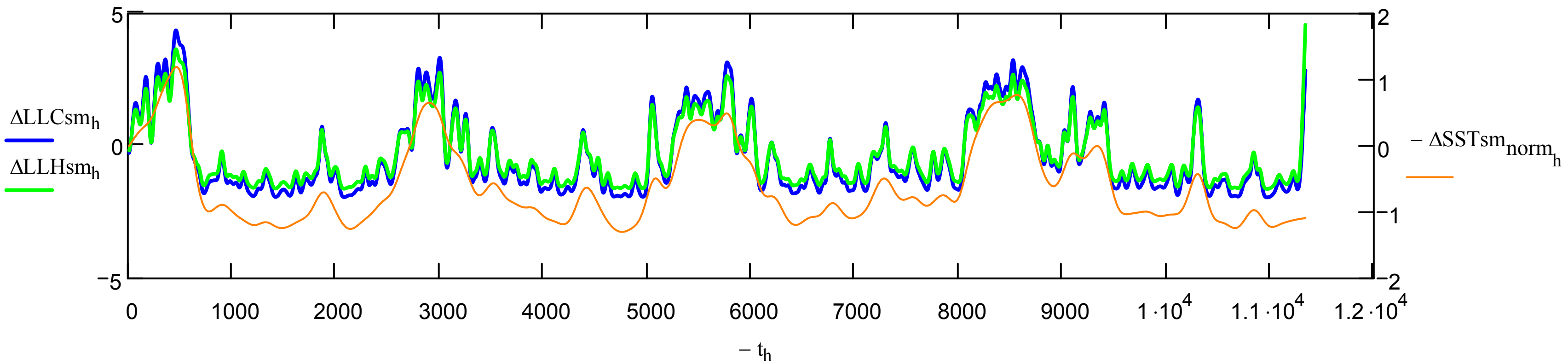
$degH = 1$

$depth0H = 30$

$t_{pres} = -15$ $\quad LLH_{pres} = -9.914$

$\Delta LLH_h := LLH_h - LLH_{pres}$ $\quad \Delta LLHsm := ksmooth(t, \Delta LLH, sm1)$

Lake-level changes relative to Present



Dead Sea specific case

$\varepsilon = \dfrac{LA}{LAmod} = 1.25 + 0.025 \cdot LL$ when LL>100 m (i.e., >-18m relative to present)

$\varepsilon = \dfrac{LA}{LAmod} = 1$ when LL<100m (i.e., =<-18m relative to present)

For parameter settings, see Rohling (2013 QSR)

$\Phi setDS := 40$

$ADSmod := 810 \cdot 10^{6}$

$\alpha := 0.5$ current evap reduction factor due to S; at near modern lake levels this is close to 0.5 (see Rohling 2013)

$evapfact := 3.25$ ratio of P over preserved P (due to evapotransp loss), set close to 3, but tune so that modern lake level = about 118 m.

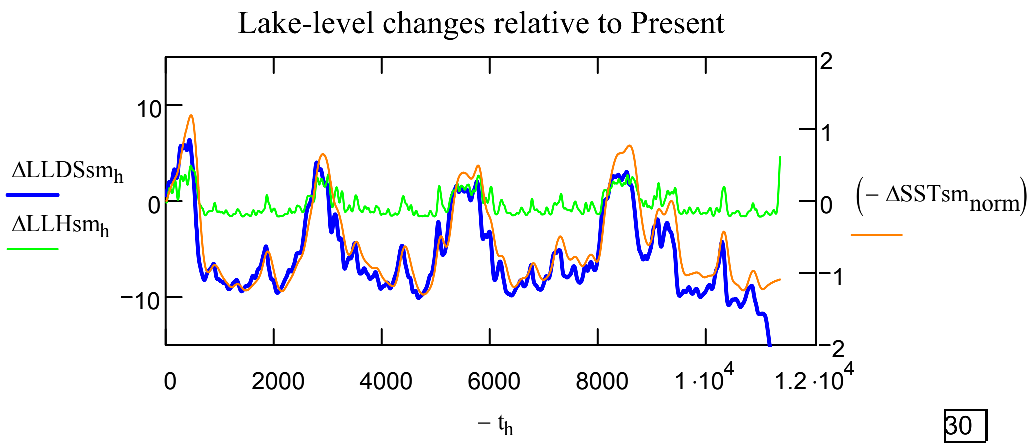
$LLDS_{pres} = 119.362$

$$
LLDS := \left|
\begin{array}{l}
LL_0 \leftarrow 0 \\
LA_0 \leftarrow ADSmod \\
CA \leftarrow (\Phi setDS - 1) \cdot LA_0 \\
\Delta LL_0 \leftarrow \dfrac{\alpha\, sumEH_0 + \alpha\, winEH_0 + \Phi setDS \cdot \dfrac{winPH_0}{evapfact}}{1000} \cdot dt \\
\text{for } p \in 1 .. pres \\
\quad \left|
\begin{array}{l}
LA_p \leftarrow \left|
\begin{array}{ll}
(1.0 + 0.025 \cdot \Delta LL_{p-1}) \cdot LA_{p-1} & \text{if } LL_{p-1} \geq 10 \\
LA_{p-1} & \text{otherwise}
\end{array}\right. \\
\Phi_p \leftarrow \dfrac{CA + 1}{LA_p} \\
\Delta LL_p \leftarrow \dfrac{\alpha\, sumEH_p + \alpha\, winEH_p + \Phi_p \cdot \dfrac{winPH_p}{evapfact}}{1000} \cdot dt \\
LL_p \leftarrow \Delta LL_p + LL_{p-1}
\end{array}\right. \\
LL
\end{array}\right.
$$

0.025 assumes that lake will fill catchment area linearly with changes in height (1/40)

$\Delta LLDS_h := LLDS_h - LLDS_{pres}$

$\Delta LLDSsm := ksmooth(t, \Delta LLDS, sm1)$



KEY RESULTS

Max. Aegean mixed layer cooling (deg. C):

$\min(\Delta SSTsm_{norm}) = -1.191$

$\max(\Delta SSTsm_{norm}) = 1.299$

Minimum snowline altitudes:

$\min(Salt\_Crete) = 923$ $\min(Salt\_Lev) = 769$

Apparent amount effects: permil per 100 mm

$me\Delta C = -0.141$ $me\Delta H = -0.5$ meΔH should be at around -0.5 according to Bar-Matthews

Control year Precip values:

$annPCone = 504.918$ $annPHone = 504.99$

Mean annual O-isotope values for different types of year:

$ann\delta PC0 = -6.04$ $ann\delta PH0 = -5.351$

$ann\delta PCone = -6.172$ $ann\delta PHone = -5.658$ NB., "one" is the control scenario!

$ann\delta PCthree = -6.398$ $ann\delta PHthree = -6.453$

$ann\delta PCfive = -6.639$ $ann\delta PHfive = -7.097$

$OUT1 := augment(t, NDDC2ksm, \Delta SSTsm_{norm}, \Delta SSSsm_{norm}, annPC, annPCsm, annEC, annECsm, annPH, annPHsm, annEH, annEHsm)$

$OUT2 := augment(t, FCsm, FHsm, ann\delta PCsm, ann\delta PHsm, \Delta LLCsm, \Delta LLHsm, y, win\delta PH, win\delta PHsm, annTaH, annTaHsm, \Delta LLDSsm)$

$OUT3 := augment(winTaHsm_{norm}, win\delta PHsm_{norm}, win\delta PHcarbsm_{norm}, \Delta\delta swsm_{norm})$

$OUTall := augment(OUT1, OUT2, OUT3)$

$Params1 := augment(cool, Vinc, \Delta\delta a, sm1, sm2, annPCone, annPHone, \min(\Delta SSTsm_{norm}), \max(\Delta SSTsm_{norm}), \min(Ta0), \min(Salt\_Crete), \min(Salt\_Lev), me\Delta C, me\Delta H)$

$Params2 := augment(ann\delta PCone, ann\delta PHone, \Delta win\delta carbsm_{3k})$

$Params := augment(Params1, Params2)$

Key operator values with some notes:

Set counters to pick specific columns

$n_T \equiv 9$ $\quad cool := n_T \cdot 1$ $\quad cool = 9$

$n_V \equiv 10$ $\quad Vinc := n_V \cdot 1$ $\quad Vinc = 10$

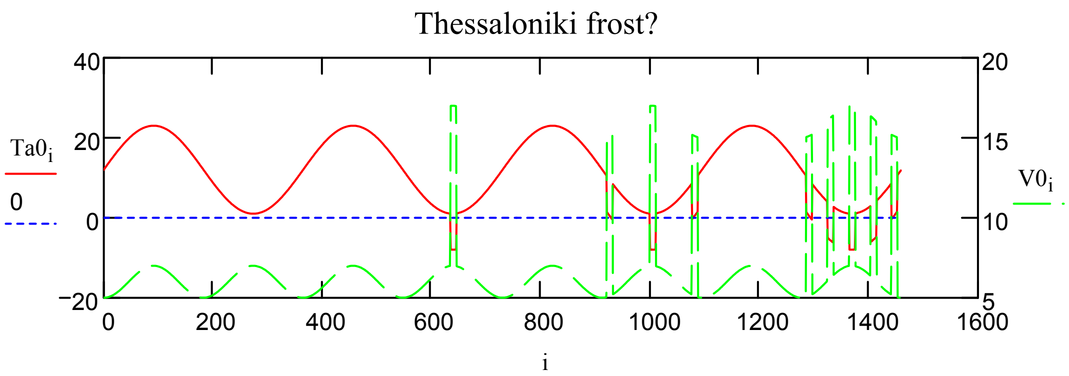
Vinc = 10 would be reasonable as Bora winds can be consistent at 17 m/s
http://www.istrianet.org/istria/geosciences/meteorology/winds/bora-adriatic.htm

$n_i \equiv 0$ $\quad \Delta\delta a := n_i \cdot -0.1$ $\quad \Delta\delta a = 0$

Then: $\max(\mathrm{win\delta PC}) - \min(\mathrm{win\delta PC}) = 0.625$

O isotope anomaly advected into region with incoming air during events (NB can be source-water effect due to NAtl 'hosing'):

Check for Thessaloniki frosts during the events:



$\min(Ta0) = -8$

References in addition to those in the main text:

Garrett, C., Outerbridge, R., Thompson, K., 1993. Inter-annual variability in Mediterranean heat budget and buoyancy fluxes. J. Climatol. 6, 900-910.
Mparsakis, S. I., 1999, Synoptic and Thermodynamic Analysis of Snowfall in the Region of Thessaloniki, Master Thesis, University of Thessaloniki (in Greek).
Wells, N.C., 1995. Surface heat fluxes in the western equatorial Pacific Ocean. Ann. Geophys., 13, 1047-1053.
Wells, N.C., 1986. The Atmosphere and Ocean: a physical introduction, 347pp. Taylor and Francis, Bristol, Pa.

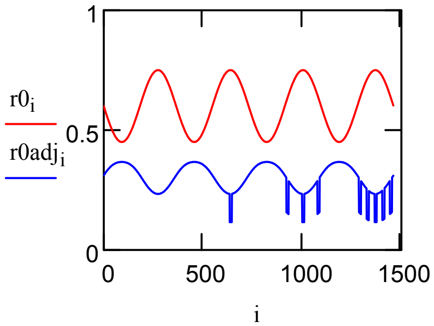
Notes to p.4:

Point in proof:

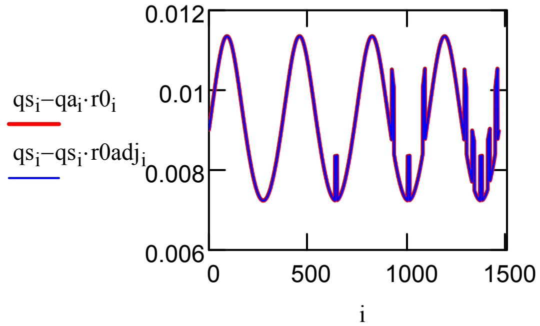
Gat et al. (2005) works through an adjusted "r" for the SST. This method gives the exact same evaporation as my method.

Gat et al. argue to calculate vapour pressure at incoming T, and then calculate how much moisture is imported (M0). Then RECALCULATE relative humidity for the SST, because that's the T at which evaporation would take place. That rel hum will be MUCH lower than the imported rel. hum. (but of course one then needs to use qs rather than qa in the evap calculation). If the model setup is correct, this will give identical answers:

$$\mathrm{pmean}_i := \mathrm{p}\cdot e^{\frac{0.5\cdot D_i}{7400}} \qquad \rho\mathrm{mean}_i := \rho\cdot e^{\frac{0.5\cdot D_i}{7400}} \qquad \mathrm{M0}_i := \rho\mathrm{mean}_i\cdot \mathrm{r0}_i\cdot \mathrm{qa}_i\cdot D_i \qquad \mathrm{r0adj}_i := \frac{\mathrm{M0}_i}{\rho\mathrm{mean}_i\cdot \mathrm{qs}_i\cdot D_i}$$



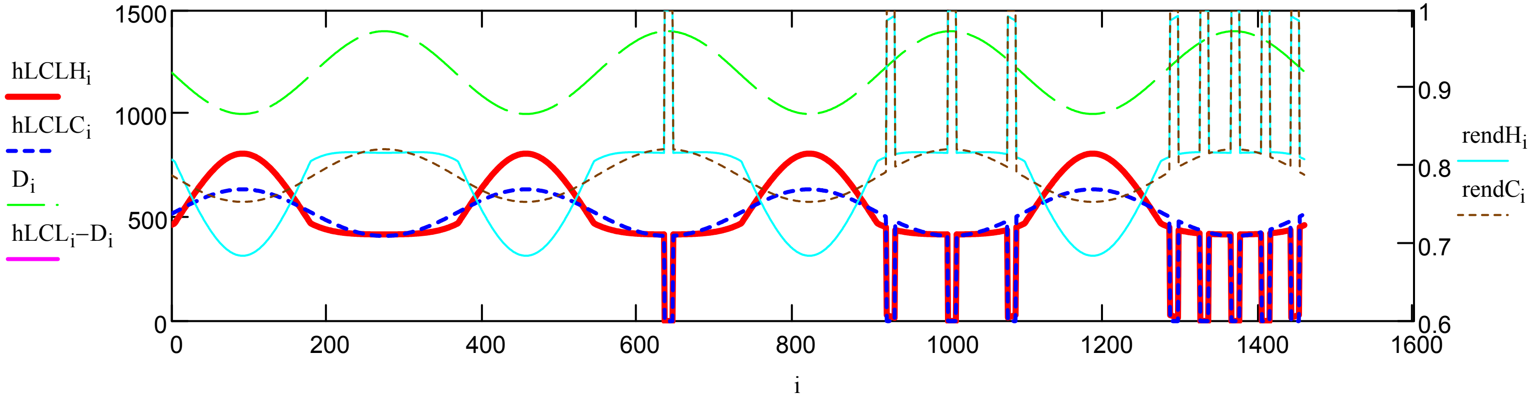
The solutions are identical:



In this extension, we approximate the cloud base, using the "Lifting Condensation Height"

$$hLCLH_i := \left(20 + \frac{TaH0_i}{5}\right) \cdot \left(100 - 100\,rendH_i\right) \qquad hLCLC_i := \left(20 + \frac{TaC0_i}{5}\right) \cdot \left(100 - 100\,rendC_i\right)$$

This uses Lawrence's formula (T in deg. C) for hLCL, the lifting condensation level height (cloud base) https://en.wikipedia.org/wiki/Lifted_condensation_level
M. G. Lawrence, "The relationship between relative humidity and the dew point temperature in moist air: A simple conversion and applications", Bull. Am. Meteorol. Soc., 86, 225-233, 2005



During events, the cloud base is seen to drop to sea level, suggesting that widespread fog/icy frost conditions mat prevail.

See also https://en.wikipedia.org/wiki/Equilibrium_level

Max height of convection is given by Equilibrium Level (or Limit of Convection). Usually is ~ the tropopause; Here we just use the observed cloud top of 5 or 6.5 km.

Note:
When air is not (yet) saturated: dry adiabatic lapse rate applies (-9.8 deg C/km)
When air is saturated, moist adiabatic lapse rate applies (closer to -6)